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STRUCTURE OF THE AKSORAN-AKZHAL FAULT ZONE IN CENTRAL KAZAKHSTAN¹

by

A. V. Luk'yanov

Folds and faults within the long-enduring Aksoran-Akhal fault zone are interrelated, commonly passing from one to another, and have a common origin. The distribution of deformations here is uneven, with comparatively undisturbed blocks among strongly disturbed ones; as a result, the entire fault zone is reminiscent of a giant tectonic breccia. The formation of individual structures is determined by their trend in relation to the zone and by the nature of their rocks.

The Aksoran-Akzhal fault zone is located in the northwestern Balkhash region, near the Kusu-Mointa watershed. It is traceable from the regional settlement in the east to the area of the Akzhal railroad station in the west. Morphologically, it is expressed in a comparatively narrow (not over 10 km wide) belt, trending almost meridionally and containing many faults, zones of crushing, shearing, recrystallization and metamorphism. It is marked by steep dips and by sharp and steep folds, absent in the rocks of the same age outside the zone. Besides its conspicuous morphology, the Aksoran-Akzhal zone is of substantial importance in the geologic history of this region, separating as it does the two geologic provinces within a long zone of Dzungari-Balkhash Hercinian geosyncline [3]. It also constitutes the northern boundary of a wide belt with numerous granite and monzonite intrusions of different ages.

This area has been adequately studied by geologists of the Ministry of Geology and Mineral Resources of the Kazakh S. S. R.; the U. S. S. R. Academy of Geology (largely by the V. S. E. G. G. G., and V. I. M. S.); the U. S. S. R. and Kazakh S. S. R. Academies of Sciences; and other organizations. The designation of the Aksoran-Akzhal zone as an independent structural unit has been generally accepted, so that there is no need to pause for that subject. Only the internal structure of this zone is considered in this article.

Participating in the structure of the Aksoran-Akzhal zone are Silurian, Devonian and Carboniferous rocks, also Hercinian intrusive

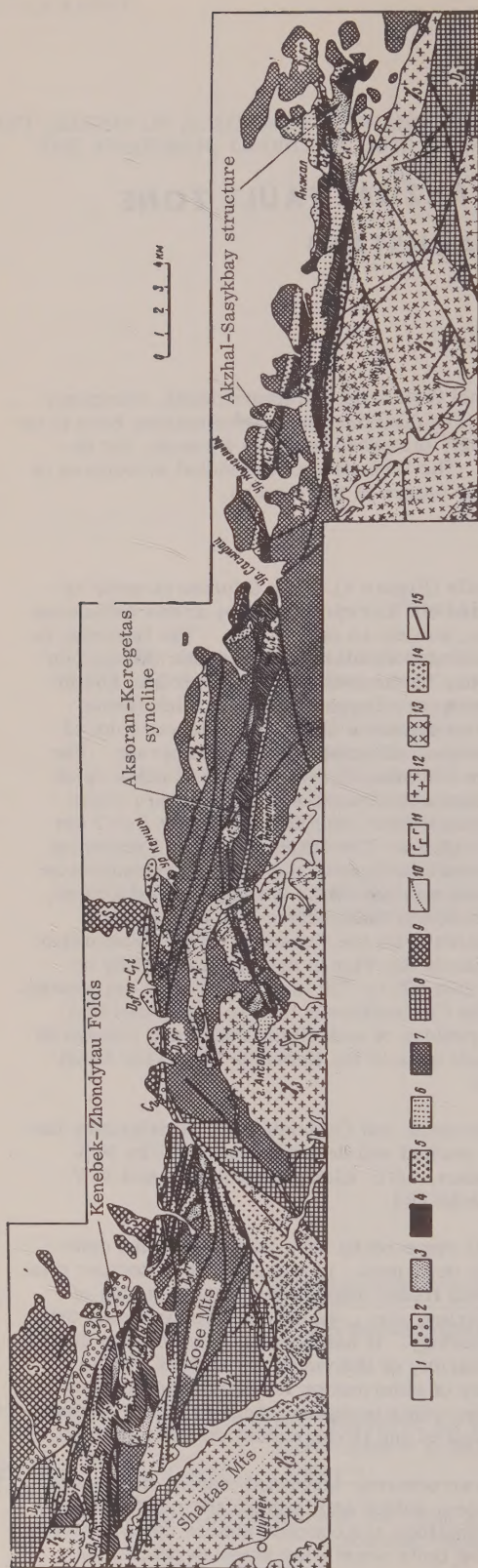
massifs (Figure 1). The Silurian is quite restricted and is represented by arenaceous rocks, subject to subdivision. The Devonian is represented by all three divisions; the Carboniferous, by the lower two. Lower and Middle Devonian are represented by a thick (about 2000 m) sequence of gray to tobacco-colored sandstone, siltstone, and conglomerate. The Upper Devonian Frasnian stage is made up of continental extrusive and sedimentary rocks common in this zone and commonly 1 to 2 km in thickness. The Upper Devonian Famennian stage and the base of the Lower Carboniferous Tournaisian constitute a single carbonate unit, 300 to 400 m thick, connected by a gradual transition with the Frasnian. The Tournaisian and partly the Viséan are represented by no less than 600 or 700 m of sandy silty carbonates. Middle Carboniferous coarse conglomerates with pebbles of underlying rocks are present in a small area of the Kenebek-Zhondyttau Mountains.

Devonian and Carboniferous stratigraphy has been studied and described in detail by N. S. Pupyshov, M. I. Aleksandrova [1], and V. V. Donskikh [6].

All these rocks have been highly and complexly deformed. In our attempt to decipher this maze of folds, homoclines, faults, zones of shattering, etc., we have studied in detail their morphology. It has been determined that despite this variety of individual structures, the intensity of deformation is distributed quite unevenly. This is typical of all Aksoran-Akzhal structures and is their most common feature.

In structures, large and small, simple and complex, folded and faulted, the most intensive deformations are concentrated in comparatively narrow belts separating areas of much less

¹Структура Аксоран-Акжал'ской зоны разломов в центральной казахстане.



intensive deformation. These narrow deformation bands are split into many branches, diverge, and come together again, in an intricate network containing segments of lesser deformation with a definite latitudinal elongation.

Thus, the lack of uniformity in the distribution of deformation is typical not of individual structures alone but also of the zone as a whole, with its lenticular blocks of slightly deformed rocks enveloped by zones of intensive deformation.

This common feature of all Aksoran-Akzhai structures, illustrated in a tectonic map (Figure 2), is reflected in assorted tectonic forms not shown on the map and calling for a special description.

MAJOR STRUCTURES

The complexity of major structures in the Aksoran-Akzhai zone reflects the complexity

FIGURE 1. Generalized geologic map of the Aksoran-Akzhai fault zone

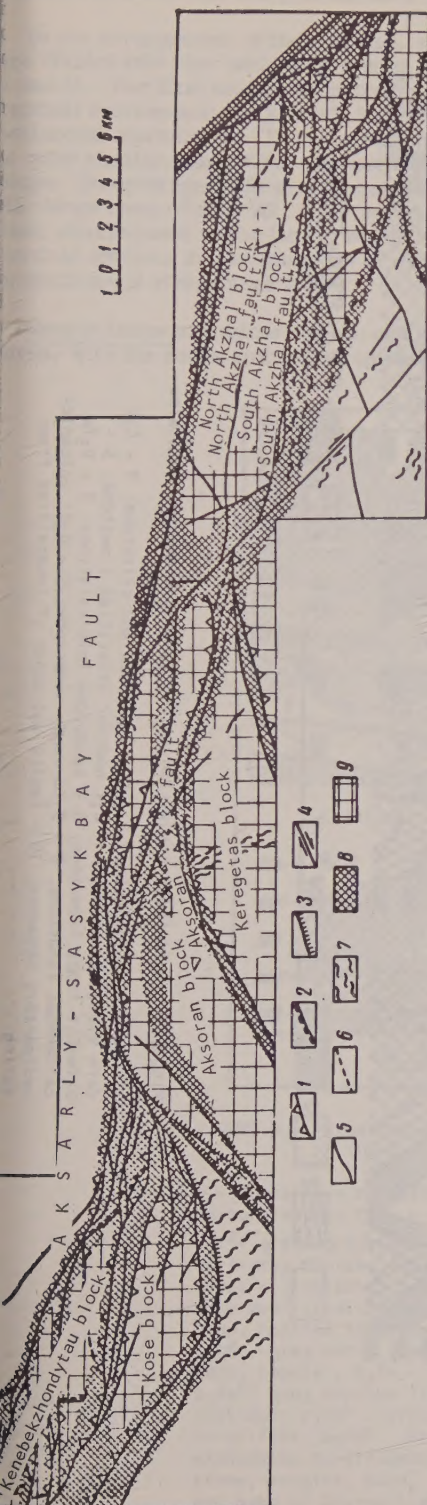
1 - Quaternary deposits - Q; 2 - Middle Carboniferous (conglomerate) - C₂; Lower Carboniferous: 3 - Sandstones - C_{1t-v}; 4 - Limestone - C_{1f}; 5 - Undifferentiated Famennian and Tournaisian (recrystallized limestone) - D_{3fm-C_{1t}}; Upper Devonian: 6 - Famennian (limestone) - D_{3fm}; 7 - Frasnian (extrusives) - D_{3fr}; 8 - Middle Devonian (sandstone, conglomerate) - D₂; 9 - Silurian - S; 10 - Acid to intermediate dikes; 11 - Gabbro; 12 - Granite - γ₃; 13 - Granite and granodiorite - γ₂; 14 - Diorite - γ₁; 15 - faults.

of deformations affecting the zone. The major structures show different deformation effects in different combinations. These effects seldom occur by themselves; consequently, clean-cut simple structures are quite rare. However, as components of complex features, they are important elements.

Simple Structural Forms

Five types of deformation are common in the Aksoran-Akzhai fault zone: 1) dip of beds; flexures; 3) break and displacement; 4) intense crushing; and 5) metamorphism going as far as complete recrystallization. Corresponding to these types are the following simple structural forms: 1) homoclines; 2) folds; 3) faults; 4) zones of intense crushing; and 5) zones of dynamic metamorphism and recrystallization.

In homoclines, the beds usually dip steeply (50 to 90°) but parallel to each other, and



maintain their mode of occurrence typical of each homocline. Homoclines are widely distributed in rocks of all ages and lithologic types. They range from a few hundred meters to 20 km long and several kilometers wide (North Akzhal and South Akzhal ridges; Keregetas Mountains, etc.). Large homoclines are complicated by occasional minor folds and faults. Areas of homoclinal rocks are ordinarily defined by faults; locally, they change to folded areas (as in the east end of South Akzhal ridge).

Folds are considerably more restricted than the homoclines. They are associated largely with Famennian and Tournaisian limestone and sandstone, and are considerably less common in Devonian sandstone and pyroclastics. Simple folds, not complicated by minor faults, have a limb span of 1 to 300 m, mostly 30 to 50 m. The vast majority of folds are asymmetrical and angular, with an extremely sharp bend at the axis.

The asymmetry of folds is expressed in their flexure and step-like structure, with a synclinal flexure always associated with an anticline.

FIGURE 2. Structure of the Aksoran-Akzhal zone

1 - normal and lateral faults; 2 - thrusts; 3 - faults with intensive shearing and metamorphism; 4 - lateral faults; 5 - assorted faults; 6 - axes of principal folds; 7 - zones of dikes; 8 - zones of intensive deformation; 9 - comparatively slightly deformed blocks.

The two mutually opposite flexures and the corresponding limbs form a single major structure. The short limb of a fold lies between the flexures, with long parallel limbs lying beyond them.

The two long limbs appear to form a single homocline whose uniform dip is disturbed in the interval between the mutually parallel hinges of the adjacent flexures. Thus, these folds differ from flexures in that they deform homoclinical rather than horizontal beds. The most intensive deformations are concentrated in the hinges and are expressed in sharp bends without breaks. The short or "disturbed" limb between the hinges has an orientation quite different from the monocline - flat, steep, commonly vertical, and even overturned. In many instances, the structure of folds becomes more complicated along its trend; faults appear in the hinges, along with the bending. In places, the nascent fault with such a hinge grows gradually to take in the "disturbed" limb, turning it to a more or less wide zone of crushing and finally completely displacing all elements of folding from such a "flexure-like" structure. Folds change gradually to faults

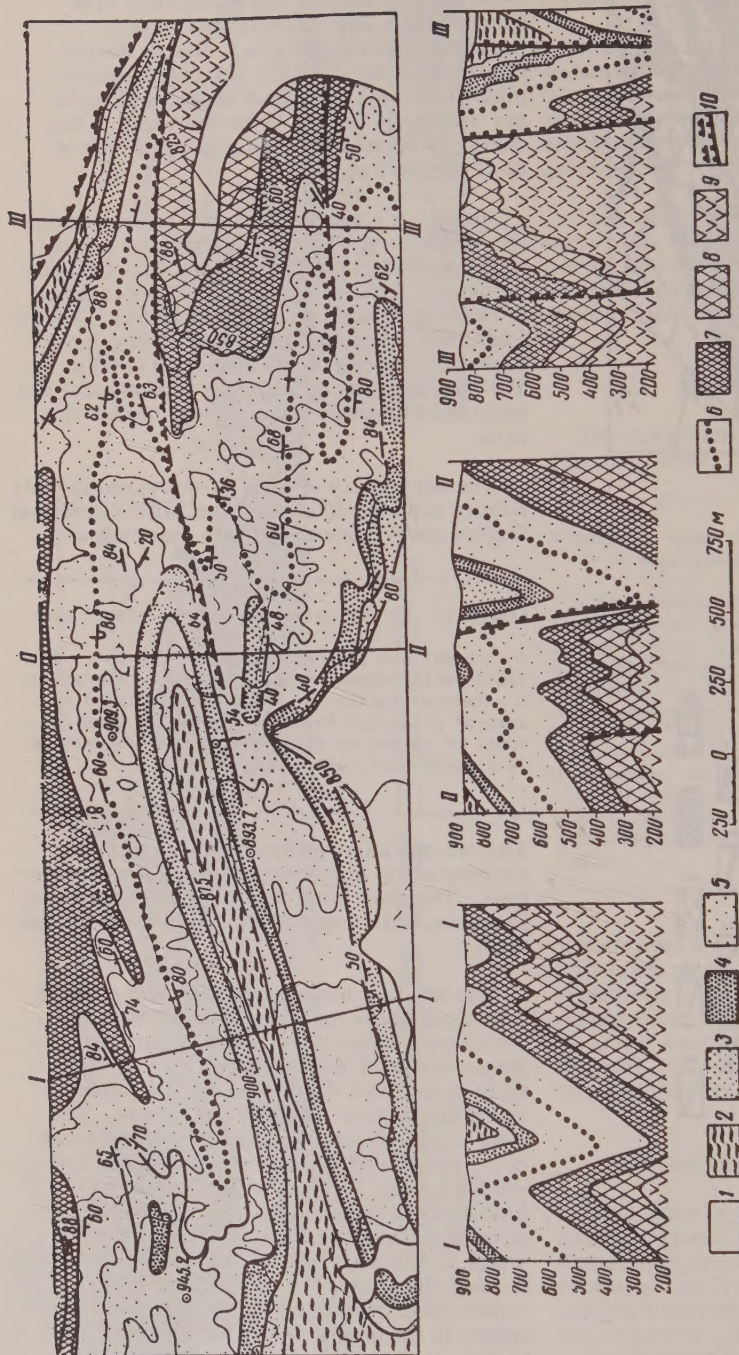


FIGURE 3. Fold-fault transition in the Kenebek-Zhondytau Mountains

1 - Quaternary deposits; 2 - C_{1t} sandstone and siltstone; 3 - C_{1t} dark slaty limestone; 4 - C_{1t} nodular stratified limestone; 5 - D_{3fm}-C_{1t} lumpy limestone; 6 - sandstone intercalations; 7 - D_{3fm} calcareous sandstone; 8 - D_{3fm} sandy limestone; 9 - D_{3fr} volcanic-sedimentary sequence; 10 - faults and zones of intensified fracturing

(Figure 3). Folds complicated by faults are common in limestone and pyroclastics.

In the arrangement of their hinges, the folds are divided into flat- and steep-hinged (Figures 3 and 4). The first originated probably in vertical movements; the second, in lateral horizontal movements. The morphology of both is quite similar, except for the position of the hinge. Steep-hinge folds are reminiscent of the flat-hinge ones rotated 90°: their structure, in plan, corresponds to the latter; while their vertical sections are reminiscent of the plan projections of steep-hinge folds.

Simple faults are minor breaks and fractures, with the rocks split and displaced along

clean-cut planes, dipping ordinarily at least 60° and commonly vertical; their sides are crushed or separated by a thin gouge; their displacement ranges from several centimeters to 200 m and has both vertical and horizontal components. Simple faults are quite common throughout this zone, in granite, pyroclastics, sandstone, massive limestone, and somewhat less common in other limestone varieties.

Zones of intensive crushing accompany large faults. They are not uncommon in the "disturbed" limbs of folds. Their rocks have been crushed to such an extent as to obliterate their original stratification. Pure limestones are marked by a more intensive fracturing; at a blow, they fall apart into small irregular chips;

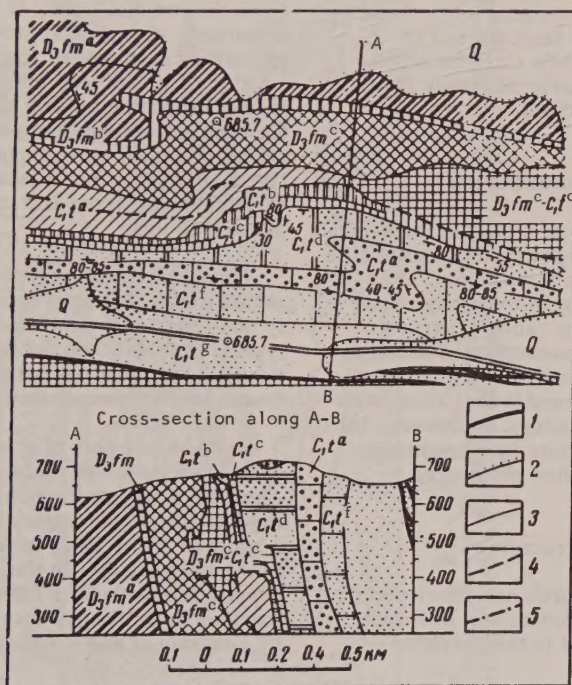


FIGURE 4. Structure of a fold with a steep hinge, in the Sultan Springs area

Q - Quaternary deposits; C1t - Tournaisian stage of Lower Carboniferous; C1t^g - green-gray calcareous sandstone; C1t^f - green-gray, coarse, clastic, sandy limestone; C1t^e - dark-gray, thick, tabular clastic limestone; C1t^d - gray clastic limestone, medium to thin, tabular; C1t^c - pinkish-gray well stratified, thin, tabular limestone; C1t^b - dark-gray, mottled limestone, well stratified, thin, tabular; C1t^a - gray marly limestone, locally sandy, well-stratified, thin, tabular; D3fm - Famennian stage of Upper Devonian; D3fm^c - gray massive limestone, lumpy, with siliceous concretions; D3fm^b - pink-gray limestone, massive, non-stratified; D3fm^a - brown-gray limestone, argillaceous, arenaceous to siliceous; D3fm^c - C1t^c - light-gray limestone, massive, pure, locally replacing top of Famennian and base of Tournaisian limestones. Contacts: 1 - fault; 2 - unconformable; 3 - normal, with gradual transition; 4 - very vague; 5 - sandstone intercalations.

such chips often form tectonic breccias cemented with white calcite. Argillaceous and siliceous limestones turn to calcareous shales or to lump breccias of limestone lumps and small lenses ground together. The sandstones are schistose; the extrusives are broken into thin slate, by innumerable fractures oriented along the fault plane. The surface of these slates is often slicken-sided and covered by a thin crust of epidote, quartz, or gouge. The conglomerates are broken up by numerous fractures which cut both the pebbles and their cement.

Zones of recrystallization and metamorphism, too, are usually associated with large faults. Recrystallized limestones have turned to marble and marble schist. West of the Zhilandytau Mountains, Frasnian extrusives have been altered to porphyroids, while Middle Devonian sandstones have been altered to quartz-mica schists. Middle Carboniferous conglomerates in the Aksarly-Sasykbay fault zone are strongly quartzitized.

Complex Structural Forms

These are a combination of simple forms. They are considerably larger than the simple forms, being measured in kilometers. In morphology, they range from folds (characterized largely by flexures) to faults (characterized by breaks and displacements). There is no sharp boundary between these two modes of deformation, in the Aksoran-Akzhai zone, with both present in every structure. However, for convenience, we differentiate this continuous series into folded and faulted structures.

1. Folded Structures

Complex folded structures are developed in the Kenebek-Zhondytau Mountains, Aksoran-Keregetas Mountains, and in the Sasykbay-Akzhai zone.

The folded nature of major structures is best represented in the Kenebek-Zhondytau Mountains of Famennian and Tournaisian carbonate rocks. Their structure is quite complex but can be generalized as two adjacent folds, an anticline and a syncline, or more precisely a single flexure.² This structure has a latitudinal trend and is traceable for about 10 km. The total width of the anticline and syncline (limb span) is about 4 km (Figures

1 and 5). In the north, south, and east, this structure is bound by faults; in the west, by a granite massif.

The flat limbs of this "flexure-like" fold, i. e., the south limb of the anticline and the north limb of the syncline are deformed in the same way, on the whole; they dip relatively gently (about 30°) to the south and are complicated by minor asymmetrical folds trending almost latitudinally. Some of these folds are slightly faulted, but the general flat dip of the limb is not disturbed. Each wing is 1 to 2 km wide, in plan.

The steep limb of the "flexure-like" fold, i. e., the south limb of the syncline and the north limb of the anticline, is quite different from the two flat limbs; it is very strongly disturbed, with typical steep dips (80 to 90°) intensive crushing and fracturing, recrystallization of limestone, and numerous fault surfaces, branching off and converging again, forming an intricate network of closely grouped tectonic lenses, throughout the limb. The entire limb is in effect a band of strongly deformed rocks, from 100 to 800 m wide, in plan and 10 km long.

Thus, the complex Kenebek-Zhondytau fold structure has features typical of faulting, closely and apparently genetically related to those typical of folds: the concentration of deformations in a narrow belt, the presence of numerous faults, and evidence of metamorphism.

Another example of a structure with well-defined elements of folding is the V-shaped Aksoran-Keregetas syncline (Figure 1), trending latitudinally and traceable for 20 km, with a width of 4 to 6 km. In the north and east, it is cut off by faults; in the south and west, by granite intrusions. Frasnian extrusives are exposed in its limbs, with the core represented by Famennian and Tournaisian carbonate rocks.

The south limb of the syncline dips north and represents a single homocline exposed in Mt. Keregetas and in the south extrusive zone, in the New Aksoran area. In the latter, it dips north-northwest 40 to 50°; in Mt. Keregetas it dips north 60 to 70°. The homocline is broken only by a comparatively few minor transverse faults.

The north limb is more complex. In the Aksoran area, its extrusives dip south to southeast 30 to 60°, while their strike changes from N - 30° - E to S - 60° - E, in smooth curves. Small but quite sharp folds are present locally. The trend of extrusives in the north syncline limb, north of Mt. Keregetas, is usually impossible to determine because of the lack of stratification; stratified sequences dip south-southeast, at 45 to 70°. Areas of flatter beds (dipping at about 15°) occur in the Samannay

²It should be noted that numerous minor folds render these structures so complex as to make it impossible to measure the general dip of individual beds. The latter reflect only local dips in minor folds with a span of a few to 100 to 150 m, just as the measurements on horizontal and vertical surfaces do not reflect the dip of a step-ladder fault zone.

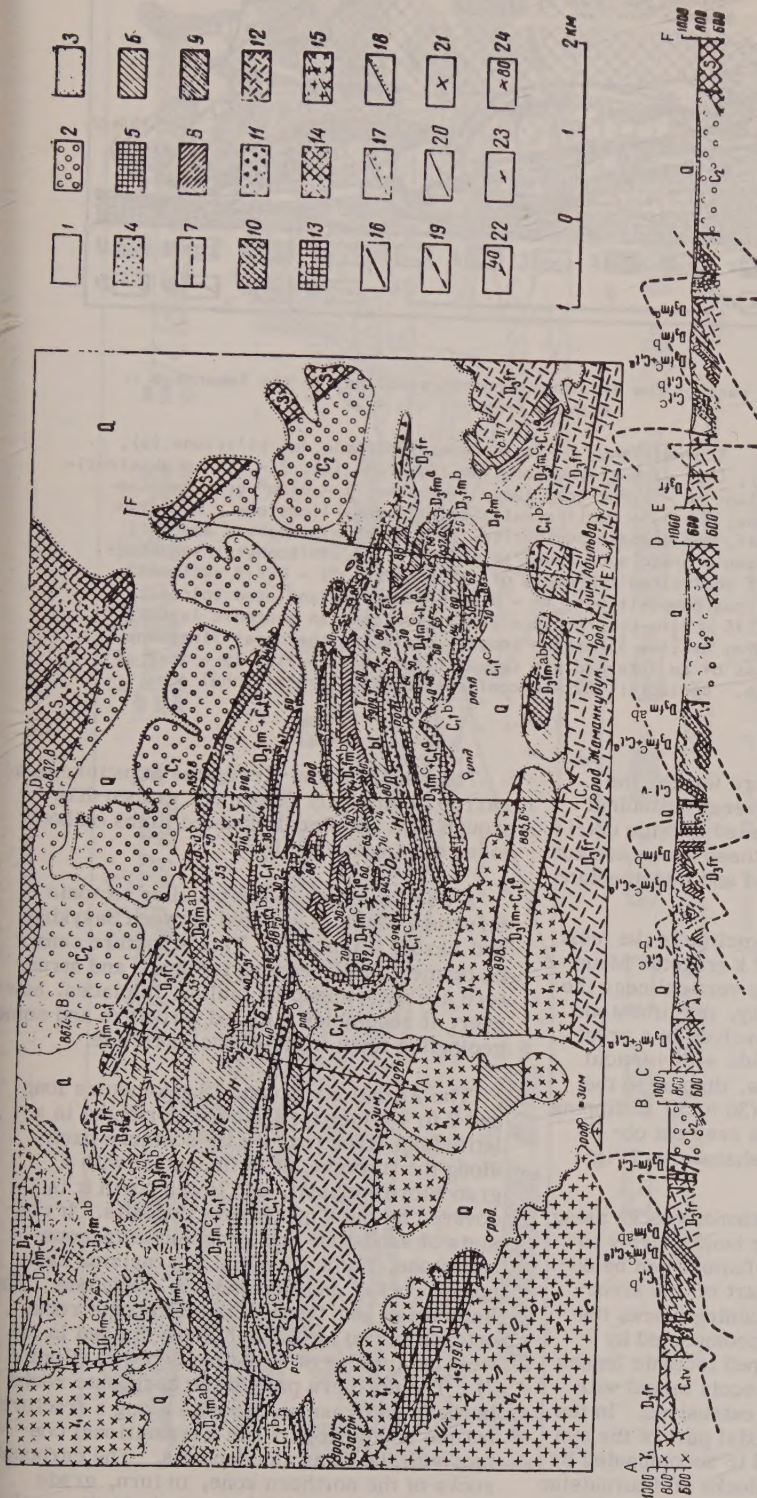


FIGURE 5. Folded structure in the Kenebek-Zhondytau Mountains

1 - Quaternary deposits; 2 - C₂ Middle Carboniferous (conglomerate). Lower Carboniferous: 3 - C₁t-v siltstone and sandstone; 4 - C₁t^c dark-gray stratified limestone; 5 - C₁tb pink stratified limestones. Undifferentiated Lower Tournaisian and Upper Famennian: 6 - D₃fm+C₁ta gray lumpy limestone; 7 - sandstone bed. Famennian stage: 8 - D₃fm^b gray sandy limestone; 9 - D₃fm^a gray calcareous sandstone; 10 - D₃fmab undifferentiated sandy limestone and calcareous sand; 11 - D₃fm-C₁t undifferentiated Tournaisian and Famennian (recrystallized limestone in tectonic lenses); 12 - D₃fr Frasnian stage (tuffaceous sandstone, tuff, extrusives; 13 - D₂ Middle Devonian (sandstone); 14 - S Silurian (shale and tuff); 15 - Granite (γ₂) and granodiorite (γ₁); 16 - faults. Contacts: 17 - intrusive; 18 - angular unconformity; 19 - sharp but without obvious unconformity; 20 - conformable, gradual transition. Manner of occurrence: 21 - horizontal; 22 - normal; 23 - vertical; 24 - overturned.

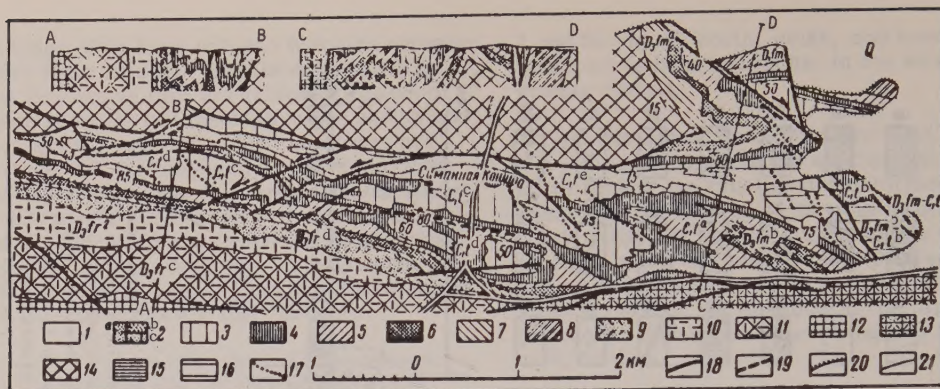


FIGURE 6. Axial part of the Aksoran-Keregetas syncline, in the Samannaya Koshara Area

1 - Quaternary deposits. Tournaisian: 2 - calcareous sandstone and siltstone (a); calcareous sandstone (b); C_{1t}^d dark-gray stratified limestone (c); 3 - C_{1t}^c pink stratified limestone. Famennian: 6 - D_3fm^b gray lumpy limestone; 7 - D_3fm^a gray argillaceous-siliceous limestone; 8 - D_3fm undifferentiated siliceous limestone. Frasnian: 9 - D_3fr^e sandstone, tuff, and diabase porphyrite; 10 - D_3fr^d albitophyre and tuff; 11 - D_3fr^c tuff and volcanic breccia of andesite porphyry and albitophyre; sandstone, breccia; 12 - D_3fr^b tuff and volcanic breccia of albitophyre; 13 - D_3fr^a tuff and volcanic breccia of various compositions; 14 - D_3fm-C_{1t} undifferentiated volcanic-sedimentary formations; 15 - D_3fm-C_{1t} recrystallized limestone in tectonic lenses; 16 - $D_3fm^b-C_{1t}^b$ light-gray massive limestone locally replacing the upper limestone interval; 17 - vein rocks of an intermediate composition; 18 - faults; 19 - zone of intensively crushed rock. Contacts: 20 - unconformable; 21 - transitional conformable.

Koshara area. Along its entire length, the north limb is disturbed by several latitudinal fractures which are accompanied by wide detrital zones. In several places these zones are cut by long, narrow strips of acid and intermediate composition rocks.

The axial portion of the syncline is the most dislocated. In the vicinity of Keregetas Mountain, the south limb is nearly vertical near the axis. Because of its plasticity, the limestone in this area is most likely involved in a great number of fine, but sharp folds with vertical and commonly inverted limbs, dipping to the north, and undulating limbs (30 to 50°), dipping to the south. These features are best observed in the Samannaya Koshara area (Figure 6).

The finely corrugated carbonate rocks along the axial part of the syncline bend sharply toward the north limb. This feature can be observed in the easternmost part of the structure (Figure 6). In the remaining places the bend is so sharp that it is accompanied by faulting with numerous trapped tectonic lenses of limestone and tectonic breccia mixed with fragments of limestone and extrusives. In the Novyy Aksoran region the axial part of the syncline is also corrugated and is accompanied by several fractures. Small blocks of Tournaisian calcareous sandstone have subsided considerably along these fractures.

Thus the Aksoran-Keregetas structure is a narrow, complexly deformed, graben-like trough with simpler dislocated limbs on both sides dipping toward it.

A third example is the peculiar tectonic structure which is situated between Sasykbay and Akzhal (Figure 7). This structure cannot be termed a fold because its shape and structure are determined by faults. In places, however, it resembles a folded structure and some geologists term it the Akzhal anticline.

The Akzhal structure, 18 kilometers long and 2 to 4 kilometers wide, is oriented in a latitudinal direction. In its southern part, along the fracture, it is in contact with a granodiorite massif, and in the north it is covered by thick quaternary deposits. It consists of sandstone and Frasnian tuffaceous sandstone, Famennian and Tournaisian carbonate rocks, and Tournaisian sandstone. The Famennian and Tournaisian carbonate rocks outcrop in two latitudinal zones divided in the west by a wedge of Frasnian deposits (Figure 7). The northern part of the carbonate rocks borders on Frasnian deposits along a fracture but the southern part has a transitional contact with the Frasnian deposits. The carbonate rocks of the northern zone, in turn, grade transitionally into Frasnian beds in the north. Individual lithologies are well defined within

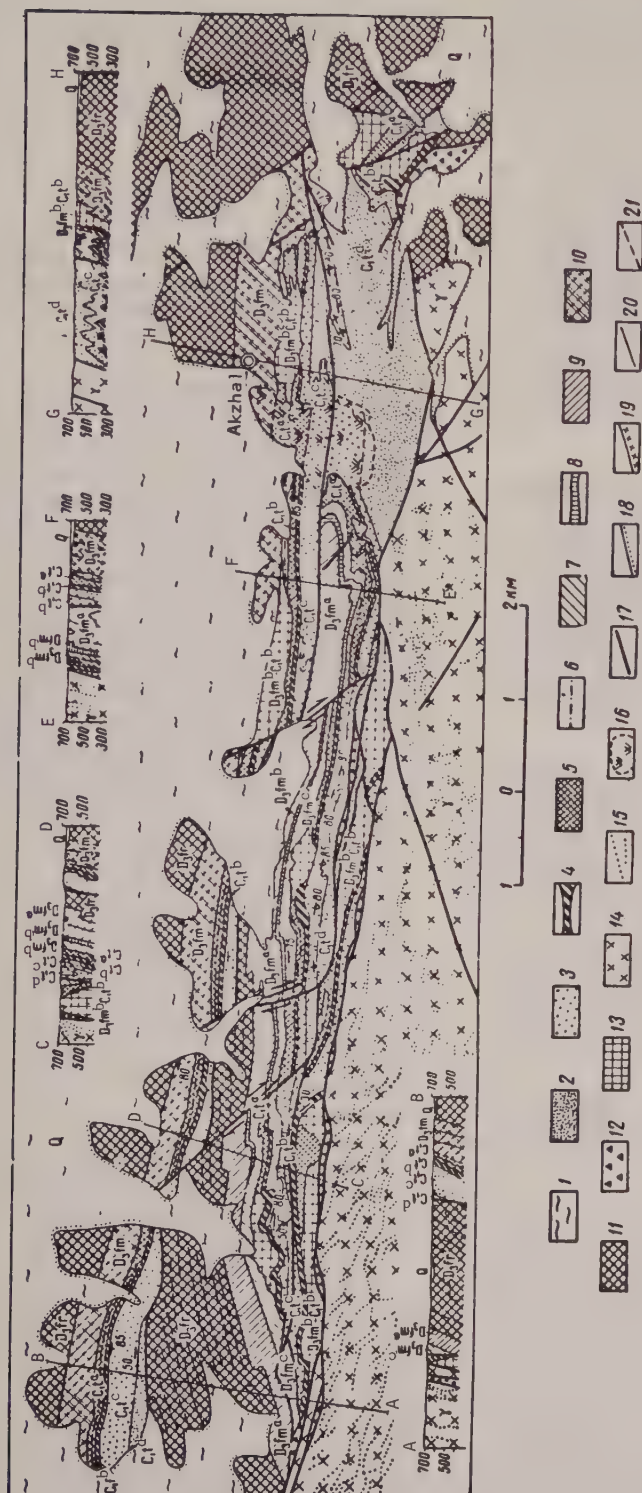


FIGURE 7. Structural map of the Akzhal area

1 - Q Quaternary deposits. Tournaisian: 2 - C1td calcareous limestones; 3 - C1tc dark-gray stratified limestone; 4 - C1tb pink stratified limestone; 5 - C1ta dark tabular marly to sandy limestone; 6 - sandstone bed. Famennian: 7 - D3fmc gray lumpy siliceous limestones; 8 - D3fmb pink to light-gray massive limestone; 9 - D3fma brown sandy and clayey limestone; 10 - D3fm undifferentiated sandy, clayey, and siliceous limestones; 11 - D3fr Frasnian stage (sandstone, tuffaceous sandstone, tuff). Undifferentiated Famennian and Tournaisian deposits: 12 - recrystallized limestone in tectonic lenses; 13 - D3fmb-C1tb light-gray massive reef limestones, locally replacing upper intervals of Famennian and lower intervals of Tournaisian limestones; 14 - granodiorite γ ; 15 - acid to intermediate dikes; 16 - skarn and hornfels; 17 - faults. Contacts: 18 - conformable; 19 - intrusive; 20 - conformable, transitional; 21 - strike of beds.

both carbonate rock zones, with the Famennian beds along the northern rim and the Tournaisian along the southern rim. Thus the entire structure is a monocline twinned by a fracture, or in other words, two monoclinical benches separated by a fracture.

The rocks in the northern bench dip to the south at angles of 70 to 90° and rarely less. In some places the beds are overturned. The strike along the entire bench is uniform and constant. In the western part of the zone the monoclinical structure of the rocks is complicated by very rare, small bench-like folds with horizontal bends and a limb span of 1 to 2 m., and vertical flexures and a limb span of 3 to 10 m. In the east (west of Kaskaaygyr deposit), Tournaisian limestones near the fault form a small synclinal fold with a limb span of 150 m.

The southern step is also steep and uniform (strike, almost longitudinal; dip, south, at 60 to 90°). It becomes considerably more complex only in the south, in a large fault zone separating sedimentary rocks from the intrusive massif. Here, the limestones and sandstones are badly crushed and displaced by faults; near Mt. Temirtas, they form a long and narrow synclinal fold, 250 to 300 m wide, accompanied by numerous smaller folds and slightly recumbent to the east.

This recumbent position of beds, locally becoming overturned, up to 70°, has been observed along the entire southern belt of limestone outcrops. Elsewhere in this southern step, not affected by the fault, the general homoclinal beds are disturbed only by a few transverse faults and small flexures with steep hinges and a limb span from a few to 100 m. Drag folds are present near a fault which cuts off the step in the north. These folds can be seen in Frasnian tuffaceous sandstone at the Zhyngelda point and in limestones of the Mt. Kaskaaygyr area.

Thus the Akzhal structure is essentially a fault doubling up the homoclinal rocks, with a suggestion of folding created by a slight bending of beds against the fault.

We see then that the deformation is distributed unevenly in large fold structures of the Aksoran-Akzhal zone, with relatively undisturbed rocks occupying most of each individual structure. Predominant here are such simple forms as homoclines and small folds. At the same time, each of these structures is marked by a belt or zone of intensively disturbed rocks. Such a zone either coincides with the axial plane of the fold, as is the case in the Aksoran-Keretas syncline, or corresponds to its steep limb, as in the Kenebek-Zhondyttau Mountains. This zone is characterized, besides the minor folding, by such phenomena as crushing,

schistosity, recrystallization, and breaks in continuity. The deformation is particularly intensive in narrow zones which then become quite similar to zones of crushing in major faults; in that event, all features typical of folding become inconspicuous and are replaced by faulting features, as in the Akzhal "anticline".

2. Fault Structures

Major faults are prominent in the Aksoran-Akzhal zone. As a rule, they are complex in structure and are accompanied by sizable zones of crushing.

The principal and largest fault is the Aksarly-Sasykbay (Figures 1 and 2). It is well marked from the Aksarly Mountains in the west to Sasykbay in the east, as a boundary between Silurian and Devonian-Carboniferous outcrops. East of Sasykbay, the fault is covered by Tertiary and Quaternary formations, but appears to continue to the Akzhal area, passing north of the North Akzhal ridge. This fault separates two provinces with substantially different geologic histories. North of it, Silurian arenaceous and tuffaceous facies are directly overlain by Middle Carboniferous conglomerate; with Devonian and Lower Carboniferous deposits missing. South of the fault, on the other hand, the arenaceous Silurian beds are missing, with a complete Devonian - Lower Carboniferous section resting directly on the Ordovician (and possibly on some Silurian) limestones.

The above-mentioned Middle Carboniferous conglomerates are exposed in a narrow band along the Aksarly-Sasykbay fault; their origin apparently was related to sharp relief formed at that time in connection with that fault. Judging from the distribution of Tertiary patches, that fault was responsible also for Tertiary valleys. Intrusive phenomena, widely developed south of the fault, are almost nonexistent north of it.

Thus, the Aksarly-Sasykbay fault separates two essentially different provinces, and the movement along it recurred for a long time (if any event, from the Silurian to the Tertiary).

Structurally, this is a zone of extremely intensive crushing, a few tens to 400 m wide, the locus of the main displacements. Often present among its crushed rocks are tectonic lenses of limestone, conglomerate, less commonly tuffaceous sandstone. They, too, are intensively crushed. The limestones are usually recrystallized and altered to massive marbles with traces of schistosity. Significantly, these tectonic limestone lenses occur not only in the immediate vicinity of carbonate outcrops but also at a considerable distance

way. For instance, 6 km north of Mt. Aksoran, Silurian and Frasnian rocks in the Aksarly-Sasykbay fault zone have lenses of Famennian limestone caught among them, 5 to 10 km away from the nearest point of its intersection with the carbonate body. This suggests great lateral displacements.

Branching off the main fault zone, both to the south and to the north, all along it, are auxiliary faults almost parallel to it. They are quite conspicuous in the Kenebek-Zhondytau area (Figures 1, 2, and 5). Here also are the North and South Akzhal, the Aksoran, and Zhilandytau faults (Figures 1, 2, and 7). They all, like the main Aksarly-Sasykbay fault, are accompanied by more or less wide zones of crushing, with tectonic lenses ground in them, of recrystallized limestones, strongly silicified segments, etc. However, unlike the main fault, they do not separate geologically different areas but lie wholly within one of them. They are second-order structures, not affecting the geology of the Sarysu-Mointa watershed in the way the Aksarly-Sasykbay fault does, and are only the companions of the first-order structures. Still, these second-order faults are important factors in the complex tectonics of this zone. It is their branching off and convergence that breaks it up into differentially displaced segments. The fairly wide belt affected by them, along the main feature, constitutes a peculiar structure known as the Aksoran-Akzhal fault zone. We shall consider some of these second-order structures.

The South Akzhal fault is traceable from Akzhal settlement to Sasykbay point (Figures 1 and 7), on a longitudinal trend separating monoclinally resting Devonian-Carboniferous limestones and sandstones from a monolithic granodiorite massif. This fault is accompanied by a wide zone of crushing with a number of tectonic lenses. The largest of them, with a maximum width of 1 km, is traceable for 15 km, from the Tkenekty springs west to the area of elevation 535.5 m. This lens consists of limestones often showing intensive fracturing and recrystallization and dividing the South Akzhal fault into two branches. This and other tectonic lenses of the South Akzhal fault are made up of highly recrystallized marble limestones; they range in width from a few meters to a few tens of meters and are up to 1.5 km long. They are separated by narrow (3 to 4 m) streaks of intensively crushed, brecciated to schistose rocks, strongly weathered at the surface.

Intruded in this fault zone, in the west, is a narrow body of fine-grained granite with a complex composition. The crushing effect is seen in a granodiorite massif to the south, as a broad (about 2 km) zone intruded by numerous dikes, parallel to the fault zone (Figures 1 and 7).

Fault planes in the South Akzhal zones are steep, apparently dipping south. The true magnitude of lateral displacements is difficult to establish because their direction is unknown. The data on hand suggest merely that this displacement is no less than 1.5 or 2.0 km, and probably much more.

Similar to this in structure are the Aksoran, South Kenebek, and a number of other faults (Figures 1, 5, and 6) trending latitudinally. They, too, are marked by comparatively wide zones of crushing and tectonic lenses. Well exposed in the Aksoran fault, near Mt. Keregetas, are tectonic breccias of limestone and extrusives. Also present here, north of the Aksoran fault, are meridionally trending complex acid to intermediate intrusives. Similar intrusive tongues accompany the South Kenebek fault.

The North Akzhal fault (Figures 2 and 8) has a different structure. It trends sublatitudinally from Akzhal settlement to point Sasykbay, has a steeply dipping surface, and a lateral displacement of at least 1.5 km in the western part, and no less than 0.2 km in the Akzhal area. Unlike the South Akzhal fault, the displacement occurred here along a single clean-cut surface not accompanied by an appreciable zone of crushing.

Most of the minor northwesterly faults have a similar structure. They show a distinct right-lateral displacement in the Akzhal area (from the displacement of vertical beds and from striation on slickensides, dipping north at an average angle of 10°). The magnitude of displacement along these faults does not exceed 300 m.

Present in the Samannaya Koshara area are similar left-lateral faults, trending to the northeast (Figure 7). In a number of instances, they cross and displace the fold axes. Most of these faults, both northwesterly and northeasterly, are short, subsequent to the formation of principal structures of the Aksoran-Akzhal zone, as demonstrated by their dike systems which cut all geologic formations and structures of the Aksoran-Akzhal zone. The largest of the northwesterly faults were active for a long time; one of them, in the Akzhal area, constitutes the eastern boundary of the entire Aksoran-Akzhal zone. Another such fault, at the Sasykbay point, is the boundary between the Aksoran and Akzhal segments of this zone and appreciably affects the distribution of Famennian and Tournaisian facies.

Thrust faults are another type developed in the Aksoran-Akzhal zone. They are very few, occurring only in the Kenebek-Zhondytau Mountains. Here, in the southwest, Frasnian extrusives are thrust over Lower Carboniferous sandstones, in a narrow block between the

granite intrusion and the latitudinal fault (Figure 5). Its thrust plane is meridional, dipping west at 15 to 40° . The magnitude of thrust appears to be on the order of several kilometers, with a lateral displacement of about 2 km. The movement took place along a zone of highly schistose rocks, not over 10 m thick, and containing tectonic lenses of limestone. The allochthonous extrusives are highly overturned. A thrust east of the Kenebek-Zhondytau Mountains has a similar structure.

Finally, faults located between the Kenebek-Zhondytau and Zhilandytau mountains, are peculiar in structure. They trend northeast and are accompanied by zones of intensive dynamic metamorphism and recrystallization. Their extrusives have been altered to porphyroids; the sandstones to schists.

THE POSITION OF VARIOUS STRUCTURAL FORMS IN THE ZONE

As noted before, the over-all structure of the Aksoran-Akzhhal zone is determined by the deformation intensity distribution throughout its extent. The uneven nature of this distribution in its component structures is their most common feature. This phenomenon obliterates the sharp distinctions between the folded and faulted structures. It is reflected in the morphologic details of these structures and in their mutual transition. With this regularity as the basis for a tectonic classification of the zone, we obtain a scheme illustrating its block structure, as in Figure 2. It is to be noted that rocks are considerably deformed throughout the Aksoran-Akzhhal zone, both within the blocks and between them. There are almost no rocks intact. The intensity and manner of deformation is substantially different in the blocks and outside them. The homoclinal occurrence of beds, with minor folds is typical of the blocks. The interblock segments are marked by faults, zones of crushing, and folds with one limb torn up. The strongly deformed limbs of the above-mentioned larger folds lie completely within the interblock spaces, while the normal limbs take up the adjacent blocks.

Thus, in speaking of the block structure of the Aksoran-Akzhhal zone, we merely underscore an uneven and discontinuous distribution of the intensity of deformation throughout its extent. In their turn, these blocks and the zones separating them are quite uneven in structure. Narrow zones of crushing of the second-order may be identified by a more detailed mapping within the blocks, and smaller segments of less disturbed rocks and tectonic lenses occur within the deformation concentration zones. A thorough going relationship exists between these blocks and the intensively crushed zones, so that the entire Aksoran-Akzhhal belt is reminiscent of a giant fault

breccia formed as a result of prolonged tectonic movements.

As seen from our description, the structures formed in those movements are quite diverse. Their specific forms probably depend on the nature of stresses and on the physical properties of the rocks. The first factor determines the trend of the several types of structures; the second is reflected in the dependence of structure on lithology.

Most common of the forms of intensive deformation in the Aksoran-Akzhhal zone are zones of crushing. Such zones have been described above; e. g., the South Akzhhal faults (Figure 7). In addition, the Aksoran and South Kenebek faults belong here, as well as a zone of crushing in the disturbed limb of the Kenebek-Zhondytau fold; and a sizable segment of the Aksarly-Sasykbay fault (Figures 1, 5, and 6). All these faults are developed in different rock and are almost latitudinal (due east to $S - 80^\circ - E$); i. e., parallel to the Aksoran-Akzhhal zone itself.

The zones of crushing are accompanied by development of fault breccias, tectonic lenses, and dikes, and are commonly pierced by minor elongated intrusions. Considerable vertical displacements are common, such as those in the most depressed segments of the Aksoran-Akzhhal zone: a trough along the South Akzhhal fault; the Aksoran-Keregetas syncline; a graben along the South Kenebek fault; and a lop-sided graben adjacent to the Aksarly-Sasykbay fault and filled with Middle Carboniferous conglomerates. These sunken segments are oriented along the zones of crushing, with the same strike of their beds.

All this suggest to us that these sublatitudinal zones of crushing are zones of tension, accompanied by a sinking of the adjacent blocks, by the formation of open fractures, and by the appearance of dikes and minor intrusions. Where these zones cut across limestones, the limestones have sharp drag-folds accompanied by intensive crushing. This is well illustrated by the "disturbed" limb of the Kenebek-Zhondytau flexure. Belonging to the same deformation type is the North Akzhhal fault, although it is not accompanied by a wide zone of crushing but rather a steep normal fault.

Another type is represented by much less common shear structures, with the displacement along a single clean-cut plane, mostly steep and trending to the northwest. Minor lateral faults, trending northeast, are present in the Samannaya Koshara area. Present in the Kenebek-Zhondytau mountain area are low angle thrusts along northwesterly to submeridional planes.

Finally, faults with zones of an intensive dynamic metamorphism and recrystallization are present between the Kenebek-Zhondytau and Zhilandytau mountains. They have a north-easterly trend and continue far to the south-west, beyond the Aksoran-Akzhal zone; they cut the main fault.

This regularity in the orientation of structural disturbances of different types suggests that the Aksoran-Akzhal fault system was a result of combined tensile stresses normal to it (submeridional) and of compressive stresses parallel to it (sublatitudinal).

SUMMARY

1. The Aksoran-Akzhal zone was long in developing and is closely related to shifts along major, long-active, Aksarly-Sasykbay fault.
2. The rock deformation is distributed unevenly and discontinuously throughout the zone; as a result, the structure of the zone is of a block nature, with the blocks shaped up gradually in the process of faulting.
3. Faulted and folded structures in the Aksoran-Akzhal zone are interrelated, without a sharp boundary between them because of the simultaneous effect of both types.
4. The several forms of deformation are determined by the orientation of structures and by lithology. The wide zones of shattering are mostly sublatitudinal; lateral faults are largely northwesterly; the thrusts are submeridional; and the zones of dynamic metamorphism and recrystallization trend south-west. Folds are developed mostly in carbonate rocks, within the zones of crushing.

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POSITION OF THE USINSK MANGANESE-ORE DEPOSIT IN LOWER CAMBRIAN DEPOSITS OF THE KUZNETSK ALATAU RANGE¹

by

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Students of southwestern Siberia have turned their attention, on a number of occasions, to the presence of manganese in Lower Cambrian deposits. Associated with these deposits are the Siberian Usinsk deposit (axial part of the Kuznetsk Alatau), ore showings of the Arga Range, manganese provinces in the southwestern spurs of the eastern Sayans and in the junction of the Gornaya Shoriya and Altay, and a number of smaller isolated localities.

In 1956 and 1957, we did field work in the Usinsk area, with the idea of determining the sedimentation conditions for ore-bearing deposits, their paragenetic associations and their relationship, and the place of manganese ores in them, in order to attack the problem of the origin of the ores better.

GENERAL GEOLOGY OF THE AREA

The Usinsk region is located in the central part of the Kuznetsk Alatau geosynclinal trough corresponding to a definite structural-facies zone of the Sayan-Altay geosynclinal province [13].

According to V. V. Khomentovskiy, the ancient section in this area opens with Riphean deposits and is generally as follows (reading upward):

1. The Kibras formation is exposed in the western part of the Usinsk region, outside the area of the map (Figure 1). It is represented by metamorphics: gneiss, quartzite, marble, crystalline schist, and porphyroids. Its visible thickness is 1.5 to 2 km; the age, presumably Riphean. It is connected by a gradual transition with the overlying Kondom formation.

2. The Kondom formation is represented here (along right tributaries of the Belaya

Usa) largely by extrusives with subordinate dark limestone. The extrusive members differ in both intensity of metamorphism and in composition. Often present along with the obvious metavolcanic green metamorphic schists and porphyroids, are relatively well preserved extrusives, including porphyritic tuffs of albitophyre, and other more acid compositions. Spilite type lavas are common.

By its age, the Kondom formation obviously belongs to the Aldanian Lower Cambrian, and in part to the Upper Riphean. This assertion is based on the fact that the Kondom formation changes upward, quite gradually, to a section having a fauna that resembles the upper intervals of the Kameshkovo and Bazankha Lower Cambrian units (Usinsk formation). The formation is 4 km thick.

3. The Usinsk formation, associated with which is a manganese deposit of the same name, is represented by a section of mixed volcanic and sedimentary rocks (see below). In this area the Usinsk formation is exposed in the limbs of the large, meridionally trending Sargai anticline, whose axis passes along the Sargai Range (Figure 1). Archeocyathids were collected from this formation, in both limbs of this structure. According to I. T. Zhuravleva [7], they are typical of the Bazaikha Lower Cambrian horizon.

The nature of the contact between the Usinsk and the overlying Sargai formation is not quite clear. In any event, no obvious gradual transition between them has been observed in this area. In some localities (see Figure 1), they unquestionably occur in a fault contact.

The Usinsk formation ranges widely in thickness from 1200 to more than 3500 m. These figures, however, are tentative. First, the contact of the Usinsk formation and the overlying deposits is unquestionably tectonic in some places; this alone precludes a determination of its true thickness, let alone its areal extent. Secondly, this formation has been subjected to an intensive and complex deformation which in itself hampers the determination of its thickness.

¹ Polozheniye usinskogo margantseвого mestorozhdeniya v otlozheniyakh nizhnego kembriya khrebtu kuznetskiy alatau.

The Sargai formation is represented by augite and andesite porphyrite and tuff. V. V. Khomentovskiy assigns it to the Middle Cambrian. Its visible thickness is at least 1000 m.

Thus the Riphean and Lower Cambrian of this area are represented by an unbroken series of marine deposits.

THE USINSK MANGANESE-BEARING FORMATION AND ITS CONDITIONS OF FORMATION

The lithology of this formation is quite diversified. Volcanic rocks, mainly pyroclastic, are widely distributed along with normal sediments — limestone and dolomite, and schists, predominately mica and chlorite.

Its composition is subject to rapid and regular lateral changes. Going from north to south, the



FIGURE 1. Lithologic map of the Usinsk formation

1 - Quaternary deposits; 2 - Devonian extrusives; 3 - Middle Cambrian? extrusives of the Sargai formation; 4 - 9 - Usinsk formation, Lower Cambrian; (4 - white carbonate rocks of a reef type: a - limestone; b - dolomite; 5 - dark-gray pelitomorphic carbonate rocks: a - limestone; b - dolomite; 6 - pink ferruginous limestone; 7 - gray mixed-grained tuff: a - porphyritic; b - albitophytic; 8 - alternating dark limestone, schistose tuffs; porphyroids; mica-siliceous-chlorite schist; 9 - red, mixed-grained tuffs); 10 - Kondom extrusives, Lower Cambrian; 11 - granosyenite; 12 - granodiorite and diorite porphyries; 14 - gabbro and gabbro-diabase; 15 - contacts between rock members: a - sharp; b - obscure (with elements of gradual transition); 16 - faults; 17 - zones of crushing; 18 - manganese ores; 19 - manganiferous limestone.

following zones can be differentiated by their facies, in the west limb of the Sargai syncline (Figure 1): 1) volcanic rocks, mostly pyroclastic; 2) multiple alternation of assorted schists and black fine-grained limestones; and 3) carbonate rocks.

Zone One has been traced approximately from the mouth of the Ponomarevka River to the south to the Chernyy Iyus River in the north. A brief description of this zone is given below.

1. The section begins with a carbonate rock member at its base. In the south, these are dark-gray to black limestones, fine-grained, slaty. In the north (between the mouth of Pereval'nyy Springs and the Chernyy Iyus River), they are white, reef-type limestone and dolomite with algae and archaeocyathids. Their thickness is 250 to 300 m in the south, and 700 to 1000 m in the north.

2. A member with gray, fine-grained tuff and subordinate mica-calcite, mica-quartz-chlorite and quartz schists, lenses of dark limestone, and thin lava flows. Albitophytic pyroclastics predominate in the south; porphyritic in the north. This member is 2000 to 2500 m thick, in the south, and 1200 to 1500 m in the north.

3. Mixed-grained tuff with a red ferruginous cement. Subordinate lenses of limestone are either white, marble-like, or pink, unevenly colored by iron oxides. Red tuffs are absent in the south. In the north, they attain several hundred meters in thickness.

Tuffs in this zone as a whole are litho- to crystalloclastic. Some of them consist exclusively of extrusive fragments of uniform basicity (porphyrite, albitite) while others have a mixed composition with fragments of extrusives and microgranular jasperoid rocks, obsidian fragments, numerous crystals of albitized plagioclase, and isolated grains of limestone. It is significant that all rocks represented by tuff fragments occur as original components in the Usinsk formation.

A peculiar feature of the tuff section as a whole is that its rocks, although quite similar in the composition of their clastic material, may be substantially different in structure and texture. Pyroclastic formations without any evidence of rounding or sorting are as common as those reworked by water. The latter exhibit as a rule, a degree of rounding with an appreciable amount of grains of a certain size (most commonly sand-silt grain size), and often a definite banded texture determined by an uneven enrichment of the rock in clastic material or by the grain size. Furthermore, clastic grains in water-reworked tuffs are found in a carbonate or mica-chlorite-carbonate cement, while the original tuffs are virtually cement-free. Thus, we deal here with a peculiar rock series showing evidence of both volcanic and sedimentary origin. This duality resulted from the fact that pyroclastic material was dumped directly in a marine basin where a considerable part of it was reworked and redeposited, prior to final burial. A whole series of rocks was formed in this way, with diversified degrees of rounding and sorting and with a very inconsistent grain - sedimentary cement volume ratio.

Despite such obvious vagaries of this complex, its components have this in common: they all consist of genetically similar pyroclastic material born in volcanic activity. The only difference is that, from facies to facies, their typically pyroclastic features vary in prominence.

Zone Two is a sort of transition from the above-described pyroclastic rocks to carbonate formations. This zone is well developed in the

middle course of the Belaya Usa River. It is characterized by a rapid and multiple alternation of black pelitomorphic limestone with schistose tuff, porphyroids, and fine-grained schists, chiefly mica to mica-quartz-chlorite. The proportion of fine-grained schist and limestone is greater in the south; the reverse is true for the tuff.

The tuff and fine-grained schist gradually replace each other, both vertically and laterally, by means of a whole range of transitional facies differing only in their ratios of clastic material to cement. Mineral composition of fine-grained schists free of clastic material is quite similar to that of the cement in pyroclastics.

The relationship is quite different between limestone and rocks of the tuffaceous schist series. Their contacts are always sharp, without any transition. Even where the limestones are definitely subordinate to schist and tuff, they carry virtually no clastic material. The lateral changes from schist to limestone are equally sharp.

For illustration, a bed-by-bed description of a Usinsk section interval along an exploration line, near the mouth of the Ponomarevka River (reading upward, in m), is given below.

This section shows that the Usinsk formation is represented here largely by fine-grained schist alternating with schistose tuff, extrusives, and dark slaty limestones.

Fine-grained schists are rusty-gray to greenish rocks with distinct schistosity planes occasionally stratified. Their main rock-forming minerals are chlorite and sericite, with smaller amounts of quartz, albite, and calcite, and a typical addition of organic matter, leucoxene, and iron hydroxides. The ratio of the rock-forming minerals varies from bed to bed, in a broad range, with alternate prominence of individual components.

Sericite, chlorite, quartz, and albite occur either in intergrowths of a homogeneous fine-scaled aggregate or individually, in extremely fine, distinctive layers and in similarly oriented lenses.

The calcite is strongly recrystallized and occurs in patches. Organic matter usually occurs in individual beds where it is concentrated in fine, uniformly oriented lenses. Leucoxene and iron hydroxides occur as amorphous accumulations, less commonly as a relatively evenly distributed pigment.

Tuffs and other extrusives of this zone have everywhere been altered to schists and metamorphosed to a considerably higher degree than in the north (within Zone One). There is only indirect evidence of their original composition.

1. Mica-chlorite fine-grained schist	1.0
2. Dark-gray fine-grained limestone	0.5
3. Same as (1) with distinct stratification	2.0
4. Felsitic porphyry, strongly altered	1.5
5. Same as (2)	2.0
6. Porphyroid	1.0
7. Interbedded chlorite-, quartz-mica-chlorite- to sericite schists, without sharp contacts between beds. Occasional small addition of silt-size clastic grains	7.0
8. Same as (2)	0.5
9. Mica-calcite-chlorite schist	0.5
10. Same as (1)	2.0
11. Fine-grained mica schist	6.0
12. Dark-gray limestone, fine-grained to pelitomorphic	0.5
13. Fine-grained schistose tuff	2.5
14. Dark fine-grained limestone	3.6
15. Mica-quartz fine-grained schist	3.5
16. Dark fine-grained to pelitomorphic limestone	0.8
17. Albitophytic tuff, schistose, medium-grained	0.4
18. Schistose quartz porphyry	0.5
19. Dark-gray, fine-grained slaty limestone	4.8
20. Porphyritic tuff, medium-grained, schistose	1.5
21. Similar to (19)	0.5
22. Similar to (17)	1.1

such as their relict structures and the nature of secondary minerals. Chances are, these schists have originally been volcanic rocks of various basicities: intermediate to basic extrusives and tuffs; or albitophytic and porphyritic formations. However, it is not completely clear, on the basis of composition of the source rocks, whether they were lava flows or tuffs.

Schists of intermediate to basic volcanic rocks are usually vivid green, with a distinct lepidoblastic structure and "augen" or banded texture. Their groundmass consists largely of chlorite, epidote, and zoisite scales and fine grains of secondary albite which constitute up to 30% of the rock, in some varieties. There is a large amount of disintegration products of titaniferous minerals. Sphene, actinolite, and prehnite are fairly common.

The schistosity of these rocks is determined by the orientation of elements of the fabric, chiefly chlorite, whose scales and fibers are uniformly parallel. Some varieties exhibit distinct elements of banding determined by concentrations of chlorite alternating with concentrations of albite intergrowths with epidote and zoisite.

The "augen" formed on more acid volcanics are yellow-gray. Their structural and textural features are the same as in the other rocks but their mineral composition is substantially different. The principal rock-forming minerals of their groundmass are sericite, chlorite, quartz, and albite, with the characteristic presence of iron hydroxides almost everywhere. These "augen" contain inclusions and fragments of albitized and sericitized plagioclase or quartz, and occasional amygdulose of secondary quartz crystals intergrown with albite. The dark

limestones here are structurally and texturally similar to those common in Zone Three described below.

Zone Three. In the southern part of this area, the Usinsk formation is represented largely by carbonate rocks, limestone and dolomite, with occasional lenses and intercalations of schist similar to those common in the north (Zone Two). Here, however, these schists are no longer the premanent components of the section.

The zone of carbonate rocks is traceable along the Belaya Usa lower course and along the Tumuyas River. Its total thickness in this area is 2000 to 3000 m, with the following types present:

1. Light-colored reef limestone constituting the background of this area. These are homogeneous, massive, recrystallized rocks with slaty partings and commonly carrying remains of algae and archaeocyathids. Occasionally occurring among them are lenses and intercalations of sedimentary carbonate breccias of light-colored limestone fragments, a few millimeters to 1.5 cm long, cemented by fine- to medium-grained calcite. They do not form sizable and consistent horizons and are not associated with any particular stratigraphic unit. There are no clean-cut boundaries between them and the enclosing reef limestones. Their origin is probably related to intraformational erosion.

2. Dark to almost black pelitomorphic limestones, evidently formed in inter-reef depressions. These are slaty rocks, homogeneous to finely and distinctly stratified (with individual layers 3 mm to several centimeters thick), and usually pyritic.

The black limestones are comparatively restricted, within this zone. They form a meridionally elongated body, as if enclosed by the reef ring. Its length is about 8 m, with an outcrop band 0.3 to 0.5 km wide.

The change from reef to black limestone is gradual but rapid, effected over a distance of several meters. The transition zones never exhibit any definite alternation of these two types but their contacts are marked by the appearance of extremely unevenly-colored, mottled and banded limestones of various shades of gray; the over-all intensity of coloring gradually decreases from black to white limestone. At the same time, the rock becomes more massive.

Occasionally present at the boundary between the two varieties are peculiar coarse carbonate breccias representing typically near-reef formations. They were described from the Usinsk deposit area by A. S. Mukhin who rightly associated them with submarine slumps and slides on steep slopes of inter-reef depressions.

3. Dolomites are distributed among the limestones in patches of bizarre outlines which do not coincide, as a rule, with the boundaries of individual beds. It appears that the bulk of the dolomite is secondary, formed on limestone. Significantly, dolomitization has affected largely the light-colored reef limestone, leaving the black inter-reef facies almost intact. The presence of primary dolomite cannot be ruled out but is difficult to demonstrate.

A distinctive feature of all these carbonates is the lack of any sizable amount of clastic material. Except for some sand and silt grains, the insoluble residue of these limestones is generally extremely small, constituting not over 1% in light-colored varieties, reaching 2 to 5% in dark ones. The occasional abrupt increase in the insoluble residue is associated with secondary silicification.

The organic content in these limestones also is low, and is measured in hundredths of a percent in reef varieties, and in tenths in black ones (about 0.2 to 0.5%). The deep coloring of the latter is evidently due to the presence of finely dispersed pyrite.

In the east limb of the Sargai syncline, the Usinsk formation has been obliterated to a considerable extent by an intrusion. Its remaining part is represented exclusively by the same types of carbonate rocks as those in the west.

It appears then that carbonates and volcanic rocks constitute the bulk of a motley complex which is the Usinsk formation. The carbonates are developed in the southern part of this area, while the contemporaneous volcanics occur in the north. Their lateral change is gradual, by

way of a peculiar transition zone with a rapid alternation of limestones and metavolcanics.

Two features of the physical composition of the Usinsk formation stand out. First is the obviously secondary importance of siliceous formations in the paragenesis. Only very thin intercalations are silicified in the area of volcanic rocks. In schists of the transition zone silica is present only as a component (mica-quartz-chlorite, quartz-albite-mica, and quartz-chlorite schists), and that not always. Limestones in the southern part of this area are, as a rule, free of silica. Streaks of lean siliceous ores appear occasionally within the ore field but even here they are of no particular importance.

Thus jaspers, siliceous schists, and even siliceous limestones and tuffites are not typical nor consistent members of the Usinsk formation.

Secondly, terrigenous rocks are also not typical of this formation. Carbonates in the southern part of the area are altogether devoid of clastic material. The latter undoubtedly entered the northern marine basin to some extent but never as an independent rock component. Clastic material was formed in the disintegration of newly-formed volcanic structures and is practically indistinguishable from pyroclastic material.

The two main facies of the Usinsk formation (volcanic in the north; carbonate in the south) are characterized by quite definite rock parageneses and associations. Each facies has two substantially different rock groups: 1) a paragenetic association of rocks properly characteristic of the given zone; and 2) rocks from the adjacent zone, as if wedging into the main association ("patristic" and "allophilic" of N. S. Shatskiy).

In the volcanic rocks' zone, "patristic" (native) rocks include tuffs (principal paragenetic components), lavas, fine-grained schists, and silicified varieties (secondary paragenetic components). The "allophilic" (migrant) members are limestone and dolomite.

In the southern zone, carbonate rocks are the "patristic" and metavolcanics the "allophilic" components. Reef limestones are the main paragenetic component of the first group with black pelitomorphous carbonates of the secondary. This association of "native" and "migrant" facies emphasizes once more their close relationship.

Thus the Usinsk formation is clearly differentiated into two complexes: a volcanic (north) and a carbonate (south) connected by a gradual transition. The components of each complex are paragenetically related to each other, both

laterally and vertically; consequently, they both constitute "formations" as understood by N. S. Shatskiy [15].

It appears that two such formations are identifiable within Lower Cambrian deposits of the Usinsk region: a volcanic and a carbonate. The first corresponds to a rock complex formed during a single volcanic cycle; i.e., the Kondom volcanics and the Usinsk volcanic zone. Corresponding to the carbonate formation (in this area) is the Usinsk carbonate zone (Figure 2).

What were the sedimentary conditions of this area, in Early Cambrian time? They were

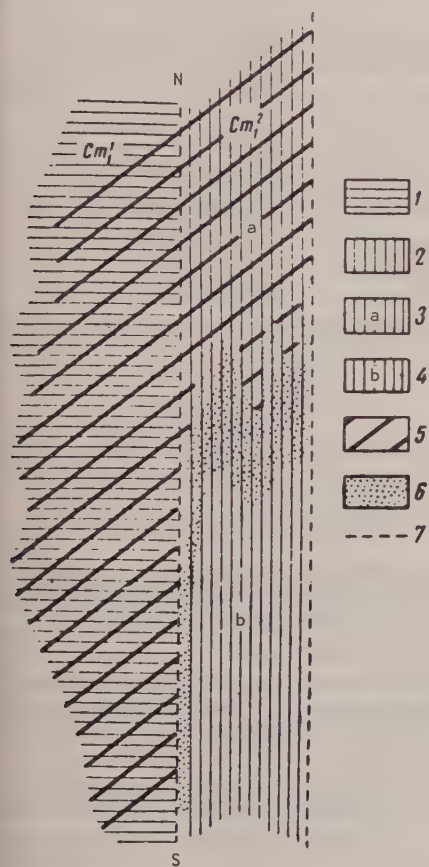


FIGURE 2. Correlation diagram for formations and members in Lower Cambrian deposits of the Usinsk area

1 - Kondom formation Cm_1^1 ; 2 - Usinsk formation Cm_1^2 ; 3 - zone of volcanic rocks in the Usa formation; 4 - zone of carbonate rocks in the Usinsk formation and the Lower Cambrian carbonate formation; 5 - Lower Cambrian volcanic formation; 6 - transition between the volcanic and carbonate formations; 7 - formation boundary.

approximately the same for the entire area; at the onset of that period, they were marked by widespread lava flows. Indeed, there is no doubt as to the marine origin of the Kondom formation. Fairly common among the extrusives are limestone lenses and intercalations, with the lavas themselves generally having a spilite aspect.

The marine basin persisted into the second half of the Early Cambrian but without the unity in sedimentation. The volcanic activity persisted only in the north, with no evidence of it in the south. The volcanism itself was of a different type: chiefly pyroclastic sequences were formed, with only local lava flows (volcanic zone of the Usinsk formation). It is to be noted that volcanism apparently was not accompanied by standard fumarole activity. The relative paucity of siliceous Usinsk rocks is evidently due to this.

That the accumulation of pyroclastic material occurred under the marine conditions, as before, is suggested by the presence of lenses and intercalations of limestone and fine-grained metashales, among the tuffs; by evidence of water reworking of the pyroclastics themselves; and by the presence of tuff varieties with a sedimentary carbonate cement.

In the second half of the Early Cambrian the formation of a thick carbonate section free of clastic material occurred in the southern part of this area. Sedimentary conditions here do not appear to have been the same throughout the area, as witness the distinctive reef and inter-reef carbonate facies. Fine black calcareous ooze was deposited in depressions among the reefs (comparatively deep and low in oxygen, judging from their fine and uniform texture and the presence of finely dispersed organic matter and a considerable amount of pyrite).

At the close of the Early Cambrian, the basin became more shallow, evidenced by the appearance of red lenses at the top of the Usinsk formation. Thus the latter was formed, on the whole, under marine conditions inherited from the Aldanian. The marine basin in the second half of the Early Cambrian was, on the whole, of an open type, although temporary volcanic islands undoubtedly existed there (in the northern part of the Usinsk region). There is no evidence of a stable dry land in the vicinity of the coast line, during the formation of the Usinsk formation, so far as this area is concerned.

This last conclusion of ours, arrived at solely on the basis of the Usinsk regional data, is corroborated by studies carried out on a larger scale. V.V. Khomentovskiy [13] points to the absence of an Early Cambrian province of long erosion, within the entire Kuznetsk Alatau eugeosynclinal zone.

A BRIEF DESCRIPTION OF THE USINSK
MANGANESE-ORE DEPOSIT

The Usinsk manganese deposit is associated with the carbonate rock zone. It is described comprehensively and in great detail in the works of geologists from the West Siberian Geological Administration: A. S. Mukhin, P. P. Ladygin, V. M. Kurshs, and others [8-9]. They all note that the ores are associated with fine-grained black limestone enriched within the ore area, bed-by-bed in leptochlorite and fine-grained silica. They also note that the latter is quite restricted. Primary ore deposits are represented by carbonate compounds of manganese, manganocalcite and rhodochrosite, which account for the bulk of the ore. Oxidized varieties or ores are relatively common only on the left bank where they are represented largely by psilomelane and to a smaller extent by vernadite and pyrolusite.

The manganese ore body has been traced for 4.6 km² along the strike; its known width is as much as 350 m. It has been found to be present at a depth of 600 m [9]. The ore is stratified in conformity with the enclosing rocks. Its strike is almost meridional, with an easterly dip (at 75 to 85°, on the average).

The deposit consists of a number of lenses elongated parallel to the strike, either chain-like or en echelon. They wedge out quite rapidly, and their contact with lateral rocks is always very sharp (A. S. Mukhin, 1940-1943).

Several faults have been identified within the ore field by the above-named students. There is a steep east-dipping fault plane along the eastern contact of the ore body and a zone of crushing along its west side. The direction of tectonic zones within the ore field almost coincides with the general trend of the formation; because of that, the faults, clearly traceable along their strike and at depth, barely truncate the ore body.

DISTRIBUTION OF MANGANESE IN THE
USINSK FORMATION

Our task was to determine the distribution of manganese in the Usinsk formation, outside the ore field; to ascertain any regularity in the change in concentration from one facies type to another; and the possibility of manganese contamination for any given horizon of the Usinsk formation. With this in mind, all the control

sections were tested for manganese.³ It has been established that just outside the ore field, manganese concentrations become normal (Clarke index) and remain so throughout the area studied. Manganese concentrations higher than normal have not been observed in any of the rock types, neither in a facies zone as a whole nor in any of its horizons.

The Clarke manganese indexes themselves vary quite regularly from one rock type to another. As shown in Table 1, manganese content in volcanic rocks is always higher than in

Table 1

Manganese Content (in %) in Principal
Rock Types of the Usinsk Formation

Rock types	Mn content
Reef limestone	0.03 to over 0.01
Inter-reef limestone	0.03 to over 0.01
Dolomite	0.01 to 0.05, less commonly over 0.01
Porphyroids	0.05 to 0.09
Albitophytic tuff	0.08 to 0.16
Porphyritoids	0.2 to 0.3

sedimentary rocks. In extrusive rocks, it rises with basicity; it remains steady in reef and inter-reef limestones; and is somewhat higher in dolomite compared with limestone.

Our figures are in fair agreement with data (Table 2) on the chemistry of similar formations, cited by other students [4, 10].

A correlation of figures in the two tables shows that none of the Usinsk rocks has an appreciably higher manganese content. It is especially significant that even limestones similar to the ore-bearing show no tendency for a higher manganese content, throughout the area.

Thus we conclude that the Usinsk formation as a whole is not contaminated with manganese. What then is the explanation for the high concentration in ore, in a single locality of this area (the Usinsk deposit)?

²From drilling data. In recent years, a lean ore horizon has been traced for a few additional hundreds of meters to the north, by geophysics. Geophysical data are not shown on the map (Figure 1).

³The manganese content in rocks was determined partly in the field, by V. N. Boltunov; partly in the Laboratory of Geological Institute Academy of Sciences, U.S.S.R.

Table 2

Correlation of Manganese Content in the Usinsk Formation
With Data of Other Students

Rock types	Mn content (%)	Author
Intermediate rocks	0.12	A. P. Vinogradov
Porphyrites from Cambrian deposits of Tuva	0.07 — 0.32	G. V. Pinus
Porphyroids of the Usinsk formation	0.2 — 0.3	
Albitophyre from Cambrian deposits of Tuva	0.092	G. V. Pinus
Albitophyre tuffs from the Usinsk formation	0.09 — 0.16	
Acid rocks	0.06	A. P. Vinogradov
Acid extrusives from the Cambrian of Tuva	0.046	G. V. Pinus
Porphyroids from the Usinsk formation	0.05 — 0.09	
Sedimentary rocks	0.067	A. P. Vinogradov
Shaly limestone in marine deposits of Donbas	0.04	N. M. Strakhov
Limestone from the Usinsk formation	0.01 — 0.03	

ORIGIN OF THE DEPOSIT

Most students of southwestern Siberia believe the Usinsk deposit to be a primary marine type, for the following reasons: 1) its generally stratified nature; 2) alternation of horizons enriched in ore substance, and barren rocks of an obviously marine origin; 3) the presence of stratified structures in the ores themselves; etc.

There is no unanimity of opinion as to the source of ore material. Some students believe that manganese compounds necessary for ore concentration were brought in by continental waters from a land where rocks occurred in a state of almost complete chemical decomposition [2, 3]. According to some other students, ore carrier solutions were endogenetic [12, 14, 16]. We are of the opinion that the first theory is in obvious disagreement with facts.

A land source of ore compounds implies two alternatives: the ore accumulation was preceded by continental conditions, in which case ore deposits can be anticipated at the base of transgressive series; or else stable land areas existed in the immediate vicinity of ore sites, during the deposition of ore-bearing beds.

Neither one of these alternatives was realized in the Usinsk deposit which is associated with the middle horizons of a continuous marine series, without evidence of any sizable erosion at the base of the ore horizon, or, for that matter, in the entire Usinsk section.

Evidence for the ore accumulation in the vicinity of a coast line is also lacking. Carbonate rocks of this area are virtually free of clastic material; and the Usinsk formation, as a whole, is not known for its terrigenous sediments.

As already pointed out, large land areas are unknown not only in the immediate vicinity of the deposit but in the entire area in question, at the time of Usinsk deposition. There were only the small volcanic islands and island ridges, in the northern part of the region, the most remote from this deposit. Those volcanic structures were made up of newly-formed rocks; they did not last long, having been subjected to recurrent submergence and emergence. It is hardly conceivable that rocks, under such mobile conditions, could have undergone a long chemical weathering to produce enough manganese for a major deposit. Thus, the concept

of an influx of manganese to this marine basin, from a land mass, is not corroborated by paleogeography.

The second hypothesis holds that fumarole emanations were the most probable source of manganese. The formation of extrusive-sedimentary manganese deposits is quite feasible. There are many deposits whose volcanic origin is unquestionable. In the Soviet Union, such are the east Bashkirian Ural deposits [12] and a group of Devonian-Carboniferous deposits in central Kazakhstan [16].

In the area under study, submarine volcanism was active during the deposition of the ore-bearing sequence; in other words, conditions favorable for mineralization existed.

In describing the east Bashkirian manganese deposits, N. P. Kheraskov set forth two possible variants of the volcanic-sedimentary hypothesis of their origin.

In the first, volcanic activity brings about a general contamination of bottom water by manganese. The position of a concentration site is determined not by the source position but by such physiographic factors as the depth of the basin, its bottom relief, and the strength and direction of the currents. Under this environment, the ore-bearing sediments are regionally consistent and are marked by a comparatively low manganese content.

In the second variant, high concentrations prevail in the immediate vicinity of a source; such deposits are crowded together.

A specific feature of the Usinsk deposit is that it lies beyond the area affected by the main volcanic center. Carbonate formations are predominant, with siliceous rocks definitely subordinate (lean siliceous tuff) and volcanics altogether missing.

It is impossible that the Usinsk ores were formed in bottom sinks, beyond the area affected by the volcanic center. These sinks represented settling vats; i. e., the concentration of ores was determined by their position rather than by that of the source of ore-carrier solutions.

Undoubtedly, the inter-reef depressions where fine calcareous ooze did settle, also presented suitable traps for manganese suspensions. This explains the position of the Usinsk ores within the reef ring and their association with pelitomorphous limestone.

It also follows that if the deposition of fine-grained limestone and manganese compounds occurred under the same conditions, their association must be regular, so that these black limestones along the periphery of a

volcanic area must show an excess manganese content, however slight. This does not mean that all such bottom depressions with black limestone were favorable for large manganese concentrations. It means rather a general tendency for a higher manganese content in deposits of a certain type; it also means their regional, even if slight, contamination.

However, the data on hand completely contradict such an assumption. The systematic tests of the Usinsk deposits mentioned previously clearly demonstrate the absence of any excess of manganese in black limestone. The manganese content of the black limestone is exactly the same as in light-colored reef limestone, everywhere corresponding to the Clarke index.

Thus, it is hardly reasonable to assume a general contamination of this marine basin with manganese, and a localization of ore bodies as determined by physiographic factors alone.

It is more probable that the ores were concentrated in close proximity to their source. Such a source must be presumed to have been beyond the main Early Cambrian volcanic province, in a zone where volcanic formations are merely "allophilic" (migrant).

It is possible that faults identified within the ore field are still older; if so, they could have been the feeding channels for mineralizing solutions. In that event, the deposit area was the site of the joint effect of two factors: submarine thermal springs debouching in a bottom area favorable for accumulation of fine suspensions, in the absence of factors promoting a dispersion of newly-precipitated compounds.

It is significant that the manganese concentrations decrease abruptly, as soon as the faults (presumably the feeding channels for mineralizing solution) leave the area of inter-reef depressions with its bulk of the Usinsk ores, and enter the reef limestones in the north and the schist in the south.

It must be noted that this areal isolation of ore-carrying channels from the main volcanic center does not militate against the volcanic origin of the Usinsk deposit. Hydrothermal-fumarole activity at some distance away from the extrusive phase is common in nature and has been repeatedly noted by students of present-day volcanic provinces (S. I. Naboko, B. I. Piyp et al.).

SUMMARY

1. The Lower Cambrian section within this area carries two distinctive marine formations: a volcanic and a carbonate, connected by a gradual transition.

2. The Usinsk manganese deposit is associated with carbonate formation but is located near its junction with the volcanic.

3. Ore components necessary for the formation of this deposit could not have been brought in from a land mass, for the simple reason that no such land existed in the Usinsk region, in Early Cambrian time.

4. The mineralizing solutions were endo-genetic.

5. The ore source was away from the main Early Cambrian volcanic center, although obviously genetically related to it.

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RHYTHMIC ZONATION IN CERTAIN GRANITOID BODIES¹

by

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INTRODUCTION

In the study of the Dzhiba molybdenum-tungsten deposit in the western Transbaykal region, in 1948-1950 and of the Shalgiya molybdenum deposit in central Kazakhstan, in 1952-1953, the author's attention was drawn to peculiar sinuous quartz "veinlets" in granitoids. Subsequently, the author has repeatedly observed similar "veinlets" and similar formations in granitoid bodies of the Sel'tey, Sartas, Zhangiz, Tas-Bulak, and other deposits in central Kazakhstan; Tyrny-Auz and Karobi in the Caucasus; and in samples from the Bugday deposit in the Transbaykal region.

A more careful study of these "veinlets" has revealed the presence of unusual and even mysterious features quite common and recurring in areas thousands of kilometers apart.

These sinuous "veinlets" have been described only in Soviet literature where they were first mentioned in 1939-1940. They were discovered in decite dikes and leucocratic granite porphyries from the area of Karobi [6] and Tyrny-Auz [5] deposits. The authors of these publications believed that the "veinlets" had been formed by emanations penetrating the sinuous fractures in a still plastic magma, at early stages of hardening.

Later on, similar "veinlets" were described in more detail from the Bugday [4] and Sel'tey [2] deposits and their presence was indicated for the Temirshi-Tas massif in central Kazakhstan [3]. The interpretation of these features by V. S. Kormilitsin and M. M. Manuylova was the same as that of the earlier authors, while V. S. Dmitriyevskiy and N. P. Bukovshina believed them to be metasomatic. Similar wolframite formations have been observed in the Montmins deposit, France [7], where they are explained as a result of recurrent rhythmic injections. However, all these conjectures,

based as they are on isolated examples, do not explain all the facts and miss the essence of this phenomenon.

In our own study of some years, we endeavored to determine the distribution of sinuous veinlets, areally and lithologically, as well as their morphologic, structural and mineralogic features, and their position relative to intrusive bodies, ore and other veins and veinlets, in order to determine their origin.

As a result, we have come to the conclusion that sinuous quartz "veinlets" are examples of peculiar zonation in granitoid bodies, expressed in a regular alternation of zones parallel to the contact and differing in the development of component minerals, i. e., common rock zones without any mineral grain orientation; and zones of one or more minerals, chiefly rock-forming, showing such an orientation. Accordingly, we substitute the term, "quartz zones" (or "oriented growth zones") for such earlier terms as "vermicular veinlets", "worms", "quartz bands", and "layer vein"; and we use "rock zones" for "lateral rocks".

1. Rhythmic Zonation in the Sel'tey and Other Central Kazakhstan Deposits

The Sel'tey deposit is a rhythmic deposit in a granitoid body. It is genetically related to a microgranite porphyry massif of the same name; a small oval body of about one square kilometer, with vertical contacts, which cuts an Upper Silurian hornfelsic arenoschistose section.

As seen in outcrops, the periphery of this intrusive, is cryptocrystalline within a few meters from the contact; it is locally amygdaloidal, and free of inclusions. Appearing farther away from the contacts and parallel to them are multiple alternating thin zones of microgranite porphyry and almost monomineral quartz zones grouped in "belts" which are separated by thick, up to tens of meters, intervals of the above-named original rock (Figure 1); or they are isolated thin quartz zones

¹O ritmichnoy zonal'nosti nekotorykh granitoidnykh tel.

alternating with thick rock intervals which carry here phenocrysts of quartz and feldspar. Still farther away from the contacts, in the central part of the massif, the zonation is missing, judging from microgranite porphyry blocks quarried.

These quartz zones are traceable for tens of meters, without crossing each other and without any constrictions. Their thickness ranges from less than one millimeter to 4 cm (usually 0.5 to 1.0 cm), corresponding to the size of the quartz grains. Their configuration is sinuous, more so in the upper part of the intrusive where the zonation is more intensive. Here, the amplitude of bends reaches 10 cm.

Some of the isolated quartz zones are thin and broken. Zones grouped in belts are usually thicker and more consistent.

These zones are made up of quartz grains oriented normal to the zone surface in any given segment, and of a size always larger than in the main microgranite porphyry body. Many quartz grains have their ends extended into the adjacent zone. Individual quartz grains have rhombohedral and prismatic faces.

All these zones are asymmetrical (Figures 2 and 3), due to a number of factors. Thus often intruding into the base of a zone² are bodies of a fine-grained rock (at the periphery of quartz grain) and occasional very small (fraction of a millimeter) feldspar grains forming here the semblance of a broken zone. The ends of quartz crystals are arranged at the opposite, inner side of these zones, where they are surrounded by skeletal quartz fringes which may be regarded as an induced growth of fine quartz grains from the rock zone. In addition, these zones exhibit evidence of orthotropism and a geometrical selection of grains; it should be emphasized that the crystal heads in any segment of any zone are oriented toward the center of the massif.

The proximate zones are roughly parallel and show similar configurations determined by the form of the granitoid body contact or else by the size and form of wiggles in the preceding quartz zones and in places even by the size and form of the crystal heads (Figure 4). Even more characteristic is the bending of the quartz zone bases about the phenocrysts (Figure 2).

The thinner quartz zones are often broken up while the adjacent thicker zones and rock zones are intact. Isolated rock zones carry fragments of quartz.

The rock zones are made up of microgranite porphyry. In strongly zoned segments, the latter is a massive, pink fine-grained rock with phenocrysts of quartz and feldspar (up to 0.4 cm across) accounting for 5 to 10% of the total. The groundmass grains are measured in hundredths of a millimeter. The principal rock minerals are quartz, microcline, albite-oligoclase, and biotite; with accessory monazite and secondary sericite and chlorite.

Quartz predominates in the rock as a whole and in its phenocrysts which are bipyramidal, with the same skeletal fringes along their edges as those around the crystal heads in quartz zones. In some phenocrysts, the inner zone is separated from the outer by an extremely fine-grained rock mass often carrying, along with standard phenocrysts, aggregates (2 x 10 cm) of isolated prismatic quartz grains (up to 0.3 cm). This fine-grained mass is granulated, locally micrographic.

The microcline phenocrysts are prismatic, locally with a "sand-clock" structure, apparently determined by an uneven distribution of sodium in pyramid faces. They contain numerous dots and shreds of perthite, also irregular bodies of albite replacing the microcline. Plagioclase of the fine-grained mass is idiomorphic with relation to microcline and quartz, with its grains carrying occasional mirmekite growths. Biotite occurs in rare scales, idiomorphic with relation to grains of quartz and microcline; it is allotriomorphic on plagioclase, as if corroding the latter by penetrating into fractures in its grains; it is locally replaced by chlorite. Monazite occurs in isolated grains, idiomorphic with relation to quartz and allotriomorphic to microcline.

No difference has been observed in the rock composition of different zones; the groundmass texture varies from micropegmatitic to microgranitic. Locally, this change is regular, reflected in the size of microgranite porphyry grains near the outer contacts of zones, about 50% larger than those near the inner contacts. In addition, graphic growth of minerals has been observed near the outer contacts, which is not true for their inner contacts. Thus, we have here a rhythmic repetition of consecutively changing zones (from the contact to the center of the massif): micrographite porphyry with an obscure micrographic texture of the groundmass — finer-grained microgranite porphyry — quartz zone — back to microgranite porphyry with a graphic structure, and so on, in the same order (Figure 5).

These rhythmically-zoned segments of the massif are cut (without being displaced) by a few thin, equally rhythmically-zoned dikes of younger microgranite porphyry which carry quartz zones parallel to the contacts and closed in the interior, with the crystal heads toward

²By "base" or "outer surface" we mean the surface toward the contact; the "inner surface" is toward the center of the massif.

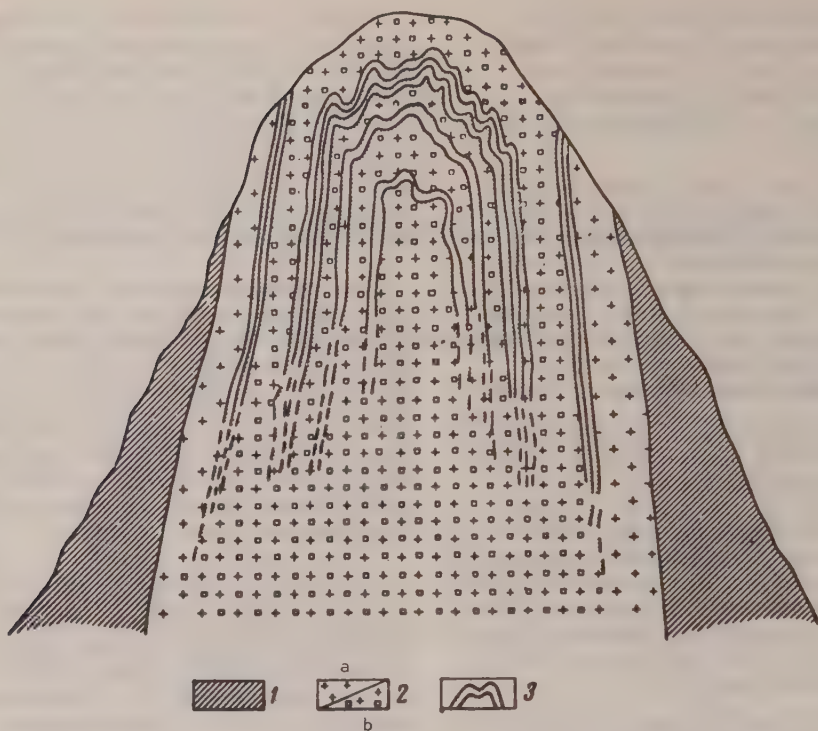


FIGURE 1. Distribution diagram for belts of quartz zones in a massif with a rhythmically zoned structure

1 - hornfels; 2 - microgranite porphyry: a - without phenocrysts; b - with phenocrysts; 3 - belts of quartz zones.

the middle of the dike (Figure 6). Quartz zones in such thin dikes are few and narrow.

As noted before, a tungsten-molybdenum deposit is genetically related to this massif. The following system of veinlets (listed in the order of their formation) is present here: 1) non-mineralized quartz; 2) quartz with molybdenite; 3) quartz with wolframite; 4) quartz with pyrite; 5) coxcomb quartz. The ore veins are associated with zones of quartz-mica greisen and occur in the microgranite porphyry massif itself and only partly beyond it.

Present in the microgranite porphyry are quartzic segments, although without any evidence of quartzitization or other alterations in the microgranite porphyry, in the vicinity of these quartzitic segments.

There are many places where the rhythmically zoned massif segments, particularly the quartz zones, are cut (without displacement) by veins corresponding to the several mineralization stages; such veins as well as the zones of these altered rocks are never sinuous, asymmetric, nor do they display any other features typical of the quartz zones.

Similar cases of rhythmic zonation in granitoid bodies are known from other central Kazakhstan ore deposits. Unlike the rhythmically zoned Sel'tey microgranite porphyry massif, those in the Shalgiya, Sartas, and Zhangiz ore fields were formed at later stages of the large Munglu multiphase granite intrusion to which they are genetically related through their occurrence at its contacts with the enclosing ultrabasic and basic rocks. That intrusion is accompanied by a complex network of veins and dikes.

These deposits are represented chiefly by veins whose mineral composition and sequences are similar to those in the Sel'tey. Their ore veins were formed after the rhythmically zoned bodies but prior to dikes of granite porphyry, diorite porphyry, diorite porphyrite and gabbro-diabase. They were followed by dikes of coxcomb quartz with fluorite and copper, lead, and zinc sulfides. Post-magmatic processes — silicification sericitization, and orthoclasonation are widely developed in the microgranite porphyries.

Here, too, rhythmic zonation is present on in granite; however, despite all the diversity in mineral composition and texture of these

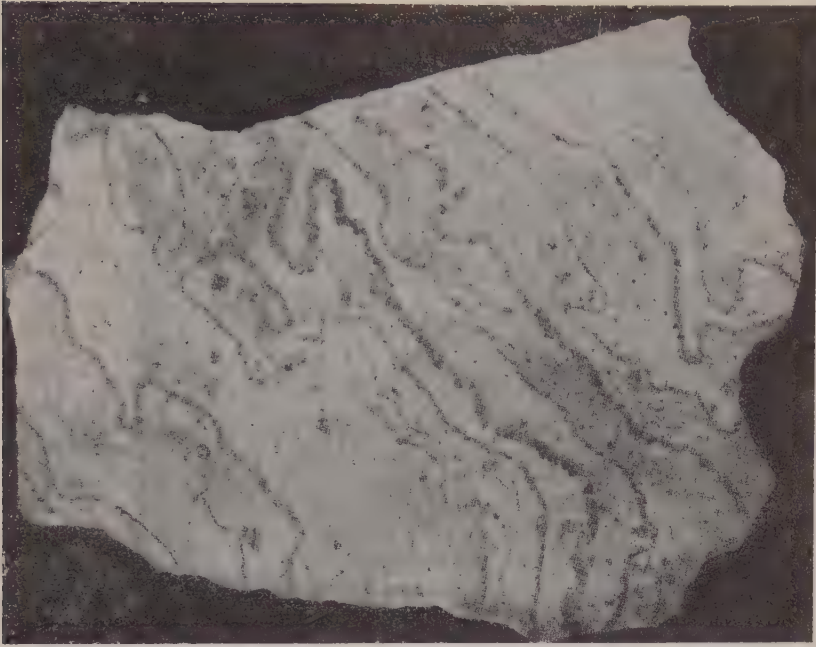


FIGURE 2. Alternation of quartz zones (dark gray) and zones of microgranite porphyry (light colored)

Note the asymmetry of quartz zones and their bases bending about the phenocrysts in rock. Thicker rock zones carry fragments of thin quartz zones. Specimen, 4/5 natural size.

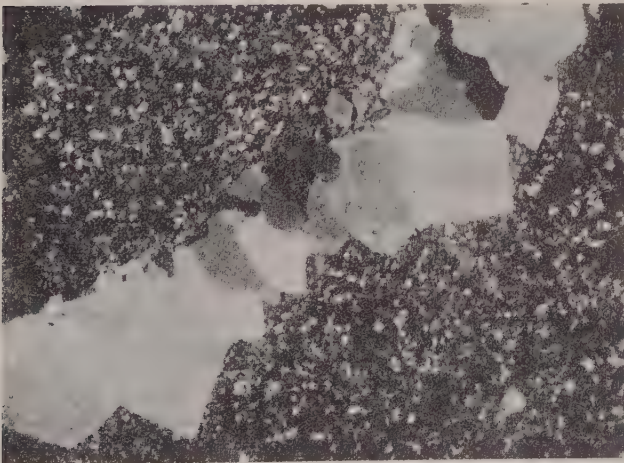


FIGURE 3. Asymmetric structure of a quartz zone, due to the difference in size and form of its quartz grains

The "heads" of these grains are idiomorphic, uniformly oriented. Thin section 5060-3/55. Magnification, 16X, with analyzer.

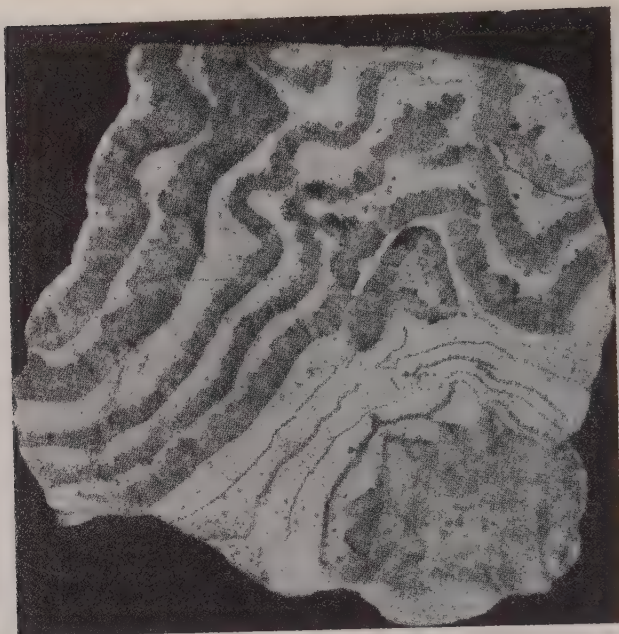


FIGURE 4. Wiggles in the rock zones (light gray) and those in quartz (gray) as a function of size and form of the crystal ends in an earlier (lower) quartz zone. Specimen, 3/4 natural size.

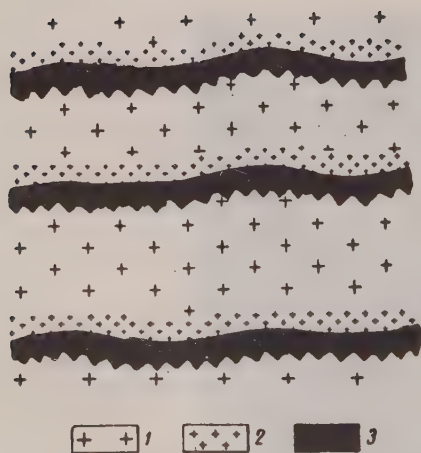


FIGURE 5. Diagram of regular alternation of quartz zones and microgranite porphyry zones of various structures

1 - microgranite porphyry with an obscure graphic structure of the groundmass; 2 - microgranite porphyry, finer-grained, without graphic structure; 3 - quartz.



FIGURE 6. Rhythmically zoned microgranite dike cutting zones of quartz and microgranite porphyry

1 - microgranite porphyry; 2 - quartz zone; 3 - microgranite zone; 4 - quartz zone. Sketch of a specimen. Magnification, 2.5X.

granites, the zonation occurs in porphyritic leucocratic rocks, microgranite porphyries, formed before the mineralization.

In addition to these forms of zonation, there is that of albite-oligoclase and occasionally of microcline; there is also the presence in rock zones, besides the standard phenocrysts, of isolated intergrowths of prismatic quartz grains with a single head (oriented in the same way as in quartz zones). These crystal heads, too, are differentiated into fine zones of microgranite porphyry parallel to the rhombohedral faces.

A conversion of the chemical analysis data by the A. N. Zavaritskiy method has shown that the rock composition is normal, supersaturated with silica. No endocontact alteration in microgranite porphyries has been observed near the hornfelsic metashales.

It is noteworthy that rock in the middle of thin rhythmically zoned dikes of Tas-Bulak often penetrates beyond the latter and into the near-contact quartz zones.

2. Rhythmic Zonation in the Bugday Deposit, the Eastern Transbaykal Region

In Bugday, according to V. S. Kormilitzin and M. M. Manuylova [4], rhythmic zonation is present in quartz porphyry dikes occurring along a fault in coarse-grained biotite granite. These dikes presumably are the roots of extrusives. The largest among them has been traced for hundreds of kilometers; it is 80 m thick and becomes thicker with depth. Its periphery is mostly felsitic; the interior, "rhythmically banded", is represented by zones of quartz and quartz porphyry; and the central part, by coarser-grained massive quartz porphyries. The transitions between these rocks are gradual.

Quartz porphyry consists of quartz and K-feldspar, with local orthoclase and isolated grains of zircon, rutile, ore mineral, and epidote. The banded quartz porphyry is rich in quartz, as compared with quartz porphyries from the central part of the dike; the groundmass structure here, unlike other varieties, is granophyric with granular sections.

3. Rhythmic Zonation in Caucasian Deposits

The rhythmically zoned granitoid bodies in the Tynny-Auz area (Caucasus) differ from those described above in composition. For this reason, we shall consider them in more detail. Their geology has been described by S. P. Solov'yev [5], while more recent data on their age relationship with other igneous bodies are presented by G. D. Afanas'yev [1]. The

largest of these bodies, the so-called "Spider", 190 x 110 m, is associated with a zone of considerable deformation. It is traceable vertically for 140 m. It is stock-like, isometric in plan, with numerous apophyses into the enclosing hornfels. Its contacts dip east and northeast, at 50 to 70°. In addition, rhythmic zonation has been observed in some of the steeper dikes (Pervaya Ravine, etc.).

These bodies consist of white granite porphyries with transitions to quartz porphyry, granophyre, microgranite, and aplite. According to L. A. Vardanyants, their principal minerals are (in %) anorthoclase, 39.43; plagioclase No. 10-20, 26.28; and quartz, 31.54; biotite is not always present (0 to 2.14); accessory minerals (0 to 0.61): apatite, fluorite, orthite, zircon, sphene, ore minerals, and monazite; secondary: albite, muscovite, sericite, actinolite, chlorite, carbonate, and epidote. Also present are pyroxene, garnet, and basic plagioclase, suggesting an assimilation of limestone. Groundmass grains are 0.05 to 0.4 mm; phenocrysts, up to 3 or 4 mm. They are represented by quartz and plagioclase, less commonly by anorthoclase and biotite.

Plagioclase (Nos. 5-92, usually up to Nos. 30-45) is unevenly distributed, being subordinate in some segments while other segments are almost monomineral plagioclase. Acid varieties replace basic plagioclase and anorthoclase or are present in interstices. Locally, normal zonation is present in the plagioclase grains: Occasional poikilitic quartz inclusions are present. Anorthoclase carries numerous decay perthites; it occurs as a fine-grained mass occasionally forming granophyric growths with quartz; also in phenocrysts strongly corroded by quartz.

According to V. V. Lyakhovich (oral communication), the structure of leucocratic granitoids is uneven and depends on the thickness of the body and the distance from the contact. The number and size of phenocrysts, too, are variable, on the whole. The groundmass is fine-grained, approaching medium-grained in the central parts of the largest bodies; mostly hypidiomorpho-granular, often micrographic, poikilitic, and aplitic.

Quantitatively, the mineral composition of the Spider granitoids is marked by the almost complete lack of anorthoclase; it is as follows (in %): plagioclase Nos. 30-45, 38.1 to 60.3; quartz, 37.9 to 61.9; ore mineral, 0 to 1; carbonate, 0 to 1.8; biotite, 0 to 2.3; and orthite, 0 to 6.

These granitoids have a higher lime content (2.46 to 5.5%), due to the assimilation of lime-rich rocks by their magma. Their silica content is quite high (up to 88.74%), as in all other Tynny-Auz leucocratic granitoids.

The contact effect of these granitoids is variable. Only biotite, muscovite, and occasionally garnet appear in biotite hornfels, while limestones are altered to garnet-pyroxene skarns, often with molybdenite inclusions.

Rhythmic zonation appears only in upper parts of granitoid bodies, as has been definitely demonstrated in the numerous mining works. V. V. Lyakhovich's observations have shown that quartz zones are not distributed throughout the largest bodies, such as the Spider and Pervaya Ravine, but rather in near-contact segments where they are conformable with the strike and dip of the contact. The work of that author brings to light another very interesting fact: rhythmic zonation formerly known only from granitoids of the Spider and Pervaya Ravine (coarser grained and with a better expressed porphyritic texture) is present also in other and smaller bodies formed by finer-grained rocks and not as porphyritic; however, these zones are not as well expressed here, and are visible mostly under the microscope.

According to V. V. Lyakhovich, quartz zones never cut the phenocrysts but stop near them; in places they are rosary-like, consisting of disjoined "drop-like" quartz inclusions.

It should be noted that grains in the Tyrny-Auz quartz zones are not of the same size. Larger grains are located usually on the bends of the wiggles, with their bases at the same level with the adjacent smaller grains. Occasionally, the quartz zones carry small amounts of K-feldspar and apatite.

Genetically and spatially related to a rhythmically-zoned granodiorite porphyry³ dike is another Caucasian rare-metal deposit, the Karobi. According to G. I. Kharashvili (1939), this dike, lenticular in plan, occurs in granites and partly in mottled schists and is associated with a steep fault plane related to the main fault which controls the neointrusions. It is 120 m long and up to 23 m thick, becoming somewhat thicker downdip. Its contact effect is represented by a recrystallization of the granites and by the formation in them of quartz and sillimanite pockets.

This dike can be seen in a quarry worked out for tens of meters, vertically. The quarry itself is lenticular in plan, with a remnant of barren granodiorite porphyry in the shape of an overturned boat, in the middle of it. Located about the remnant was an ore zone several meters thick, presenting a rhythmically zoned

dike segment. There are no quartz zones within several decimeters from the dike contacts.

In upper parts of the dike, granodiorite porphyry was ore-bearing throughout its thickness; below that, the mineralization extended in a "belt" near the contacts. In still lower horizons, judging from material obtained in drifts and boreholes, this "belt" appears to wedge out.

In the ore "belt", granodiorite porphyry zones are 1 to 2 cm thick, seldom up to 50 cm, being on the average 1 cm (rarely up to 3 cm). Thin (less than 1 mm) quartz zones are characterized by more capricious wiggles and by frequent convergence and divergence.

At their base, these quartz zones are made up of amorphous quartz grains, finer than along the other contact where the faces of the crystal heads are well-defined as they protrude into the fine-grained rock (Figure 7). The microscope shows, however, that some of these "heads" are made up of finer quartz grains. The faces of crystal heads, like the quartz phenocrysts, are covered by an induced growth of fine-grained rock.

A distinctive feature of Karobi is that its quartz zones carry molybdenite in scales idiomorphic on quartz and arranged at an angle to the zone base. These scales increase in size, gradually, from the base to the inner zone contact, along with a gradual growth of the quartz grains, from hundredths of a millimeter to 5 mm; their number, too, is greater at the base of these zones. Individual larger molybdenite scales extend from the base of quartz zones to their inner contact and protrude into the adjacent fine-grained rock of the next zone; occasionally, they grow wider, in the same direction.

Phenocrysts within the rock, be they quartz, biotite, or plagioclase, are girded on three sides and at the same distance, by the bases of quartz zones, always with a thin fine-grained layer in between (Figure 8). Where the quartz zone comes close to quartz phenocrysts, it becomes the latter's outer zone. In that event, this segment of the quartz zone is represented by a single quartz grain, with the same optical orientation as the phenocrysts whose outline it repeats (Figure 9). The outer phenocryst zone is separated from it by a rock zone which thins here, down to a fraction of a millimeter. On the inner side of quartz zones, phenocrysts grow without being separated from the quartz zone by an intermediate rock zone (between the phenocryst and the quartz zone; see Figure 8); nor is there any bending of this quartz zone contact about phenocrysts. Zonal quartz inclusions are missing where there are no quartz zones.

³Because of their distinctive porphyritic aspect, these rocks are known in literature as dacites. Because of their obviously intrusive nature, we call them granodiorite porphyries.



FIGURE 7. Asymmetric structure of the quartz zone

The grains are smaller at its base than at the opposite contact where the idiomorphic crystal heads protrude into the rock. Both quartz and rock zones are cut by fine rock veins. Thin section 37968-1. Magnification, 8X; with analyzer.

The rock zones are light gray with a porphyritic texture. Their bulk consists of grains (0.02 to 0.05 mm) of quartz, plagioclase, anorthoclase, and biotite. Phenocrysts (up to 1.5 mm) are represented by quartz, plagioclase, and biotite. Plagioclase is zoned, varying from No. 38 in the nuclei of grains to No. 28 along their periphery; it is polysynthetically twinned. Apatite occurs as an accessory mineral. In axial parts of thicker dike segments, phenocrysts predominate over the groundmass. Apatite is occasionally present in quartz phenocrysts as it is in quartz zones and in the fine-grained rock mass. Biotite in phenocrysts is unaltered even in the immediate proximity of quartz zones.

4. Analysis of the Data Presented

It follows, first of all, from these data that the above-described quartz zones cannot be the result of either a replacement of the lateral rock nor "vermicular veinlets", as believed by earlier students. Such an assumption is contradicted by their position solely in upper parts of granitoid bodies, and parallel to the contacts; as well as by their sinuous form with reniform interior bends; by their asymmetric structure (with evidence of orthotropism in minerals); the bending of the quartz zone bases about the phenocrysts, regardless of the latter's composition; the lack of altered rock near quartz zones; etc. A determination of their age may shed some light on their origin.

All quartz zones were formed before quartz veins and veinlets, including the ore-bearing,

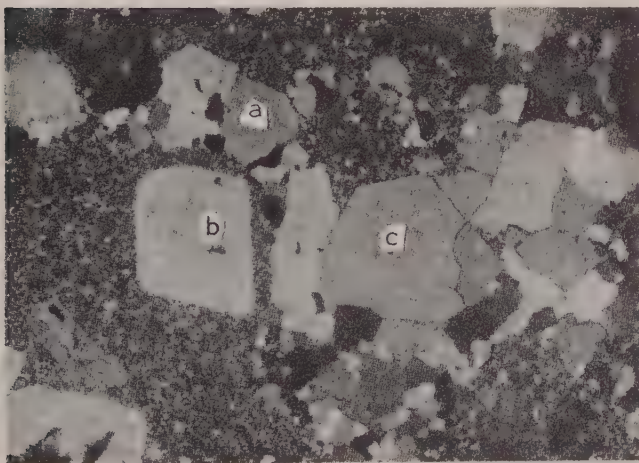


FIGURE 8. Base of quartz zone (a) bending about a quartz phenocryst (b) in rock

At the opposite contact, quartz phenocryst (c) grows into the quartz zone. Thin section 37968. Magnification, 16X; with analyzer.

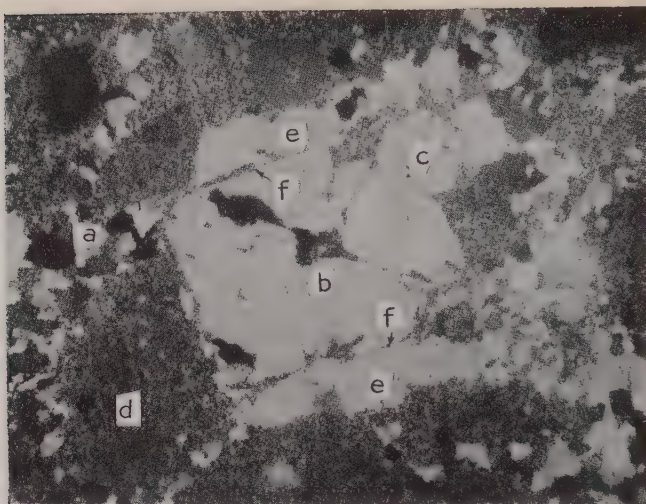


FIGURE 9. Zoned quartz phenocryst

The outer zone (a) of phenocryst was formed as the result of the induced growth of individual grains in quartz zone (c) on nucleus (b); (d) is a thin rock zone; (e) are plagioclase phenocrysts; (f) biotite phenocrysts.

but are associated with later intrusive phases of the massifs and with isolated minor intrusions cut by those veins. Neither the quartz zones nor phenocrysts are present in endo-contact zones of relatively large rhythmically zoned bodies; they occur at several decimeters or meters from the contact; and the farther away they are, the shallower is the depth of their wedging out.

Their base is bent smoothly, roughly parallel to contacts of the granitoid body and locally forms reniform loops into the latter. The quartz zones are missing in the interior of larger bodies where the porphyritic texture is less definite.

In addition, these zones are fine-grained in very fine-grained rocks and coarse-grained in fine-grained ones; they are altogether missing in medium- to coarse-grained rocks.

In very fine-grained rocks, the zones have bizarre shapes, with more wiggles and local reniform loops; in places they merge and wedge out. This relationship between the grain size in quartz zones, their form, and the structure of the "enclosing" rock is not accidental. It suggests that all these phenomena, the rock texture, the form of quartz zones, the grain size, and the very presence or absence of rhythmic zonation are possibly determined by a single factor, namely the similarity of crystallization conditions for both rock and quartz zones, in any given instance.

At the same time, such facts as the bending

of the quartz zone bases about the phenocrysts in rock, the growth of these phenocrysts into quartz zones from the side of their opposite contacts, and the bipyramidal faces of these phenocrysts, indicate that they are represented by high temperature α -quartz, while the quartz zones are made up of a lower-temperature β -quartz. This means that they are non-contemporaneous formations, with the phenocrysts being the older. It may be assumed then that the magmatic melt temperature fell below 575°C, in the time interval between the crystallization of phenocrysts and that of the quartz zones.

The relationship between quartz zones and the fine-grained "enclosing" rock is more complex. All grains in these zones, regardless of composition, have their growth oriented from the base to the interior of granitoid bodies, as witness their growth in size and the degree of idiomorphism in that direction, as well as the arrangement of their growth zones, the association of microcline grains with the bases of some essentially quartz grains, and the irregular boundary of these microcline grains with rock as against their strict idiomorphism in relation to the zone quartz. The bases of large β -quartz crystals are flat, without any induced growth of quartz grains from the bulk of the rock. That suggests that this rock zone in contact with the quartz zone base was passive in relation to it, being denser than the bulk of rock nearer to the center of granitoid body. Thus it was the inner surface of the rock zone that constituted the critical factor; otherwise, the crystals of various minerals in the oriented

growth zones would have grown from different levels and would have shown faces on all sides. Consequently, the rock zone in contact with the quartz zone base was formed prior to the latter. The crystallization of minerals within the quartz zone occurred toward the interior of the granitoid body; i. e., in the direction of the advance of the crystallization front for a given segment and for the entire granitoid body.

Let us turn now to the other contact of the same quartz zone. Here, the quartz crystal heads protrude into the rock. Like the faces of quartz phenocrysts, they show an induced growth of quartz grains in the fine-grained rock body, forming skeletal fringes with the same optical orientation as the crystal heads. It follows that these rocks, at the interior contact of quartz zones, were formed after the quartz zones.

Thus we have determined that the rock zone at the base of a quartz zone is older than the quartz zone, while the rock zone with crystal heads protruding into it from the quartz zone, is younger. The logical corollary is a non-simultaneous formation of the quartz zones themselves. Indeed, the fact that any zone, particularly a thin one, bends at a short distance from a crystal head protruding from the zone in contact with the granitoid body, shows clearly that the quartz zones themselves are not contemporaneous formations and that those nearer the granitoid body contact are the earlier, i. e., it is the form of the zone boundary, nearer the contact with the granitoid body, that determines the form of the wiggles of the adjacent granitoid and quartz zones.

Turning back to the rock zones, the above-mentioned relationship between the grain size in rock and in quartz zones, and between the form of quartz zones and the rock texture, appears to be determined by the similarity in time and conditions of formation for quartz zones with those for the fine-grained rock body. In addition, there is a local and regular change in texture across the strike of the zone: from microgranite porphyry to the quartz zone to microgranite porphyry with a graphic texture, gradually changing to a finer-grained rock of the same composition but without the graphic texture, then back to a quartz zone, etc. away from contacts with the intrusive body). This zonation in the rock itself closely follows the wiggles of quartz zones. This means that changes in the rock texture reflect those in the crystallization conditions which led to the formation of quartz zones, while the appearance of crystallization conditions sets up conditions favorable for the formation of rock zones with the original texture.

All these facts, as well as the zonal structure in individual quartz grains are adequately explained by a consecutive rhythmic deposition

of alternating rock and quartz zones, from the granitoid body contact toward its central part. A differential crystallization of the entire rock body is demonstrated also by the thin recurrent injections of the rock, locally also rhythmically zoned (Sel'tey, Shal'giya, Tas-Bulak, Bugday).

However, certain facts have been observed which call for additional explanation. For example, individual segments of quartz zones look like "chains" of crystals in the rock mass, with the crystals elongated normal to the "chain". The bases of quartz zones show a penetration by the fine-grained mass from adjacent zones, along the grain boundaries. In addition, some rock zones show inclusions of quartz zone fragments, as if cemented by a fine-grained rock mass without any evidence of deformation, as is also true for the thicker quartz zones. All these facts can be explained by a crystallization of thin quartz zones at the time when the preceding rock zones, at least in the adjacent peripheral parts, were not completely hardened, and these deformations in quartz zones could have taken place at the onset of crystallization. This does not contradict the fact that each zone, nearer the granitoid body contact, was formed earlier than the quartz zone farther from the latter; it merely refines our ideas on the status of that zone at the time of its crystallization.

On the basis of the above exposition, we can reconstruct the following sequence in the formation of these rhythmically zoned granitoid bodies: 1) the formation of an endocontact altered zone; 2) the formation of phenocrysts in an uncrystallized magmatic chamber; 3) alternating deposition of quartz and rock zones, progressing from the contact to the central part of the rock body, as a result of the change in crystallization conditions; the growth of crystals in quartz zones themselves proceeds in the same direction; and 4) repeated intrusion and crystallization, at times also rhythmic of granitoids, usually very few and thin.

According to many students (G. D. Afanas'yev, D. S. Belyankin, A. G. Gokoyev, V. S. Dmitriyevskiy, V. V. Lyakhovich), the composition of crystallizing magma itself was similar to the eutectic mixture, with a higher silica content. The presence of zonation in crestal parts of porphyritic fine- to very fine-grained rock bodies suggests a comparatively rapid crystallization. A crystallization temperature close to the eutectic point is indicated by the presence of both high- (α) and low-temperature (β) quartz, as well as by the graphic texture of rocks and the incomplete isomorphic immiscibility of K-Na-feldspars.

Under such physicochemical conditions, a reasonable explanation for rhythmic crystallization is, generally, as follows. Assuming a pre-eutectic mixture of two non-volatile

components, its cooling will be accompanied by a precipitation of the component in excess over that of the eutectic equilibrium; correspondingly, the liquid phase will be progressively impoverished in that component, until the eutectic point is reached, at which time all components will crystallize.

However, with a considerable loss of heat, crystallization of the excess component may not take place throughout the rock but will be confined to its contact with cold lateral rocks; with the temperature gradient from the center to the periphery sufficiently great, and with diffusion hampered, the following situation may arise.

A contact of the melt with a cold medium will cause the formation of a crystalline crust of component A, in excess over that of the eutectic equilibrium. With diffusion slowed down, the composition of the melt in the zone adjacent to A will approach the eutectic, and the temperature will be low enough for its crystallization; consequently, a eutectic layer will be formed. However, at some distance from the contact, the melt composition will still be the original, with an excess of A, while the temperature will be too high for its crystallization but sufficiently low for a crystallization of A; consequently, that component will again be precipitated. This will bring, once more, the composition of the adjacent melt down to the eutectic; as the corresponding isotherm progresses from the periphery to that zone it will bring about its eutectic crystallization, and so on. This line of reasoning may be accepted as a qualitative explanation of the rhythmic crystallization process in the presence of a considerable loss of heat.

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METASOMATIC ZONATION IN GREISENS ASSOCIATED WITH ALASKITE GRANITE OF THE KURAMIN RANGE (CENTRAL ASIA)¹

by

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While studying the greisens associated with alaskite granites of the Kuramin Range, our attention was attracted by their zonal structure. The sequence of zones and their mineral composition in the greisenization area are subject to the same regularities. This paper is an attempt to explain these regularities in the light of the theory of metasomatic processes, developed by D. S. Korzhinskiy.

The area of greisen development is a promising subject for a comprehensive study of that process, because of the good exposures, helped by deep mining works (up to 300 m) and of the absence of appreciable younger metasomatic and metamorphic phenomena.

Alaskite granite with which the greisens are associated forms a stock-like massif about 50 km². Their age is assumed to be, somewhat tentatively, Permian to Lower Triassic [1, 8]. Several varieties have been identified (from older to younger): even-grained alaskite granite of the main phase; porphyritic biotite granite of the supplementary phase; and several vein facies — porphyritic biotite-hornblende granite, aplitic granite, aplite, and granite porphyry.

The unaltered granite consists of quartz (32.2%), microcline (46.0%), plagioclase (albite-oligoclase No. 10-15; 18.9%), and biotite (2.9%). They all are marked by a granitic texture.

Accessory minerals in the granite are magnetite, fluorite, cyrtolite, thorite, monazite, ilmenite, fergusonite, sphene, and orthite. Qualitative relationships of rock-forming minerals mark most of this granite as alaskite.

GEOLOGY AND PETROGRAPHY OF THE GREISENS

The greisens are a series of elongated parallel bodies dipping northeast at 70 to 80°.

They are associated with a fracture system of the same trend. These greisens are from several tens of centimeters to a few hundred meters long and several centimeters to 5 m thick.

They are arranged en echelon and unevenly, more numerous in some places and less so in the others. In the areas where they are concentrated, they form closely spaced bodies presenting extensive greisenization fields. Elsewhere, individual greisen bodies are separated by stretches of unaltered granite, from one to a few meters wide.

The greisenization phenomena are most intensive at depths of 75 to 125 m; they practically disappear at 270 to 300 m.

A diagram showing the greisen structure is presented in Figure 1. A quartz-topaz-muscovite vein, often carrying microcline, wolframite, fluorite, and sulfides: pyrite, chalcopyrite, arsenopyrite, galena, sphalerite, etc., is generally present in the central part of a comparatively large greisen (75 cm and over). Metasomatically altered rocks of the greisen proper are located on either side of the vein. Their composition changes systematically from the vein walls to unaltered granites, by way of a number of metasomatic mineral zones.

Directly contacting the vein is a monomineral quartz greisen zone, 2 to 25 cm thick, consisting of isometric quartz grains, 0.1 to 0.4 mm. Occasionally present here are subordinate fine muscovite scales associated with inter- and intragranular fractures.

This is followed by a bimineral quartz-topaz greisen zone, 5 to 30 cm thick. The quartz-topaz ratio is variable, with topaz usually accounting for 20 to 25% of the total. The rock is heteroblastic, with component mineral grains isometric, 0.1 to 0.4 mm long.

Topaz forms prismatic crystals, 0.1 to 0.5 mm, seldom up to 2 mm, with occasional radial growths (Figure 2). Under the

¹Infil'tratsionnaya metsomatcheskaya zonal'nost' v greyzenakh, svyazannykh s alaskitovymi granitami kuraminskogo khrebt'a (srednyaya aziya).

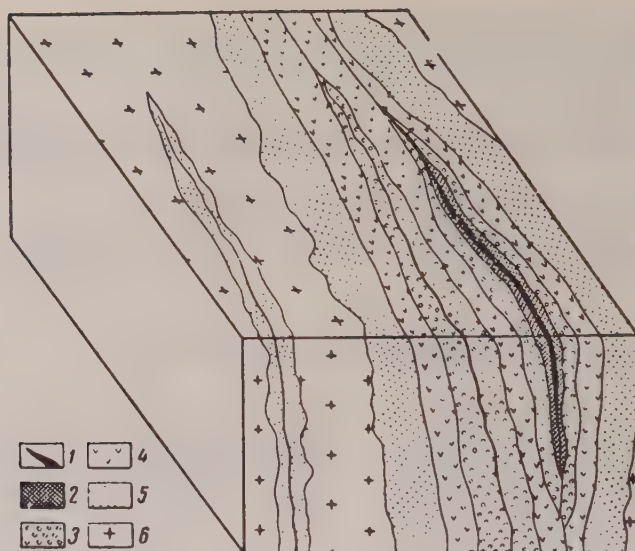


FIGURE 1. Generalized structure of a greisen

1 - quartz-topaz-muscovite vein with sulfides; 2 - quartz greisen; 3 - quartz-topaz greisen; 4 - quartz-muscovite greisen; 5 - greisenized granite; 6 - unaltered granite.

microscope, the topaz is colorless, although quite distinctive among the quartz grains, because of its high refractive index and good (001) cleavage. It is characterized by positive elongation, right extinction angle; $(+2V = 55^\circ)$; $\gamma = 1.627 \pm 0.001$; $\alpha = 1.610 \pm 0.002$; $\gamma - \alpha = 0.017$.

The chemical composition of topaz from the quartz-topaz zone is given in Table 1 (after I. V. Dubrova).

The iron in topaz is possibly explained by the presence of a small quantity of pyrite.

The boundary between the quartz and quartz-topaz zones is comparatively definite but not sharp, with a zone of a fairly rapid transition where topaz is present only in corroded relicts intensively replaced by quartz.

Adjacent to the quartz-topaz greisen is a quartz-muscovite greisen zone, from 1 to 3 m

Table 1

Chemical composition of topaz
(O. P. Ostrovskaya, Analyst)

Composition	Weight %	Molecular amount
SiO ₂	34.93	583
Al ₂ O ₃	52.34	512
Fe ₂ O ₃	0.46	30
H ₂ O ⁻	2.15	119
H ₂ O ⁻	0.70	—
F	15.20	800
Sum	105.78	—
Correction for F ₂ =0	6.38	—
Total	99.40	—

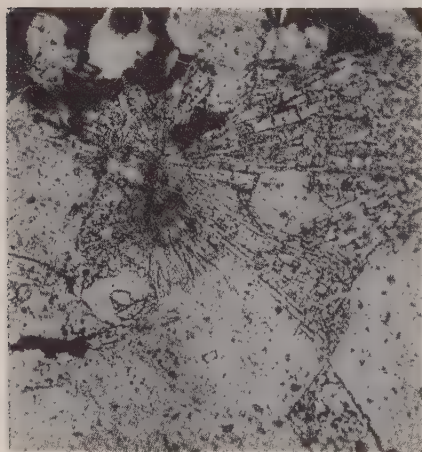


FIGURE 2. Quartz-topaz greisen

Magnification, 15X; Single Nicols.

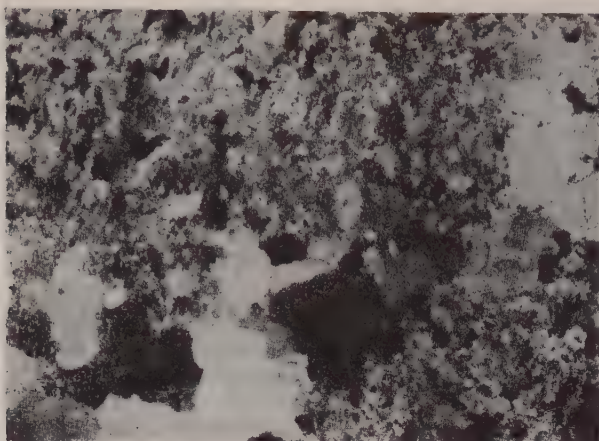


FIGURE 3. Quartz-muscovite greisen

Magnification, 15X; Nicols crossed.

thick. This is the thickest zone in all the greisens. Locally, it carries a small amount of topaz. This zone is readily identified at sight, because of its high muscovite content and the lack of feldspars. Its texture is granolepidoblastic, marked by the presence of isometric quartz grains and muscovite tablets (Figure 3). Two types of quartz have been identified: a) in isometric grains, 0.3 to 2.0 mm, with relatively rectilinear outlines; and b) in smaller grains, 0.05 to 0.2 mm, with bizarre serrate outlines. The smaller grains are closely associated with muscovite and form quartz-muscovite aggregates developed at the expense of feldspar.

Muscovite occurs either in isolated scales or in pale gray massive aggregates. Its optical properties are as follows: $(-2V) = 47^\circ$; $\gamma = 1.587 \pm 0.002$; $\beta = 1.582 \pm 0.002$; $\alpha = 1.553 \pm 0.001$; $\gamma - \alpha = 0.034$. Its chemical composition, after I. V. Dubrova, is given in Table 2.

The quartz-muscovite greisen is separated from unaltered granite by a band of greisenized granite, several tens of centimeters thick. The transition from one to the other is gradual.

The sequence of replacement reactions is fairly well expressed in greisenized granites, with two metasomatic zones identifiable in that sequence: a) quartz-muscovite-microcline, with unstable plagioclase; and b) quartz-muscovite-microcline plagioclase, characterized by stable plagioclase.

Plagioclase of the quartz-muscovite-microcline zone is partially to completely replaced by muscovite (Figure 4). Pseudomorphs of muscovite on plagioclase preserve the latter's form, so that the hypidiomorphogranular texture of the granite is preserved. It is of

interest that, along with individual plagioclase grains, muscovite replaces perthite growths of albite in microcline.

Rocks of the quartz-muscovite-microcline-plagioclase zone, in direct contact with unaltered granite, are hypidiomorphogranular and differ from unaltered granite by the albitization of their plagioclase and by instability of biotite which is altered to an aggregate of muscovite and magnetite.

We have described only the simplest structure of a greisen body, where the feeder fractures are widely spaced. Where they are closely spaced, the greisen is considerably more complex; for instance, individual zones

Table 2

Chemical composition of muscovite (O. P. Ostrovskaya, Analyst)

Composition	Weight %	Molecular amount
SiO ₂	43.94	731
TiO ₂	0.15	1
Al ₂ O ₃	33.95	332
Fe ₂ O ₃	0.77	5
MgO	0.71	17
CaO	2.36	42
Na ₂ O	1.15	18
K ₂ O	8.84	93
H ₂ O ⁻	5.84	322
H ₂ O ⁺	1.58	—
Sum	99.29	—

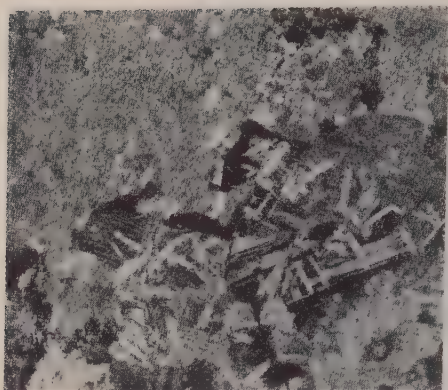


FIGURE 4. Selective replacement of plagioclase by muscovite

Magnification, 70X; Nicols crossed.

of the metasomatic section may be repeated several times. Characteristically, the lower (inner) greisen zones are always accompanied by the upper (outer) zones; there are no examples of interior zones missing; the quartz-topaz zone is always fringed by the quartz-muscovite, and the latter by the quartz-muscovite-microcline. In a number of places, the lower zones are completely missing; the quartz and quartz-topaz zone is missing in thin greisens (up to 50 cm), with the central zone represented by a quartz-muscovite zone. Veins usually are also missing in such instances.

It should be emphasized that while interior parts of greisen bodies are marked by comparatively sharp zonal boundaries, the boundaries

are very vague in the outer zones, with a very uneven distribution of differentially altered rocks. This appears to be due to a considerably lower saturation of rocks by solutions, away from the feeder channel (fracture), as well as to generally uneven porosity of the rocks. The solutions seeped through mostly along fractures, where the metasomatic alteration is the most intensive; sharp zonal boundaries arise only when the porosity and saturation of rocks are uniform.

A detailed study of greisens establishes the changes taking place in solutions, not only in space but in time as well. Thus quartz and quartz-topaz greisens exhibit muscovite scales associated as a rule with fractures (Figure 5). Replacement of topaz by muscovite occurs in places in quartz-topaz greisens.

Such phenomena are apparently related to a lowering in the acidity of greisenizing solutions at final stages of their activity. Changes in the nature of solutions, with time, such as arise in alkalinity, create additional difficulties in the study of zonation; for this reason, the nature of mineral precipitation should be closely watched. In the column below, the younger muscovite is naturally not taken into account. It appears that the formation of veins filled with quartz, topaz, microcline, muscovite, and sulfides is associated with a stage of acidity.

The formation of subsequent veins as a result of acid leaching, in connection with the decreasing acidity of solutions, is an important phase in postmagmatic processes, as emphasized by D. S. Korzhinskiy [5-7].

It is generally impossible to sample any zone

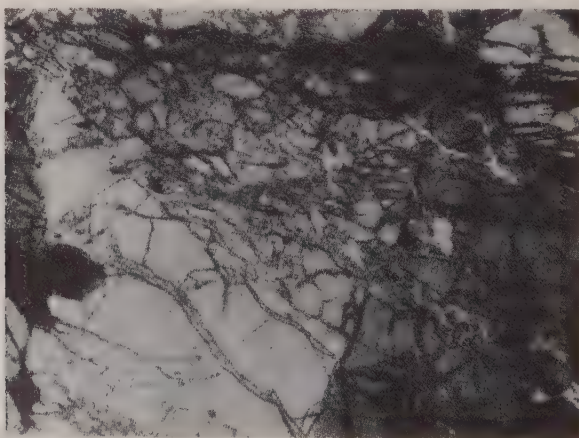


FIGURE 5. Precipitation of younger muscovite in a fracture in quartz-topaz greisen

Magnification, 70X; Nicols crossed.

for analysis, macroscopically, in such a way as to have the sample entirely free of unstable relict minerals and subsequent veins; because of that, a rigorous analysis of changes in the chemical composition from zone to zone is impossible. We have analytic data for a quartz-muscovite greisen, readily subdivided macroscopically. Comparison of the data with those for granite (Table 3) shows a greater amount of silica, compared with the unaltered granite, and a reduction in the iron, calcium, and sodium content.

ANALYSIS OF METASOMATIC COLUMN IN THE LIGHT OF THE PHASE RULE AND THE SETTING UP OF MOBILITY SERIES

According to D. S. Korzhinskiy [5], in a standard example of infiltration metasomatism, with equilibrium factors represented by the concentration of perfectly mobile components, masses of inert components, and the temperature and volume of rock, the relationship between the number of simultaneously stable minerals and that of inert components is expressed

Table 3
Chemical Analyses of Rocks

Oxides	Granite ¹		Quartz-muscovite greisen ¹		Quartz-topaz greisen ²	
	weight %	molecular amount	weight %	molecular amount	weight %	molecular amount
SiO ₂	72.76	1211	77.80	1295	75.00	1249
Fe ₂ O ₃	1.15	7	0.34	2	0.24	1
FeO	2.08	28	0.69	10	—	—
TiO ₂	0.18	1	0.18	1	0.02	—
MnO	0.05	—	0.12	3	—	—
Al ₂ O ₃	11.93	117	12.42	122	18.25	178
CaO	1.15	20	0.60	11	0.39	6
MgO	0.25	6	0.30	7	0.12	3
K ₂ O	4.88	52	4.02	42	1.46	16
Na ₂ O	2.68	44	0.35	6	0.19	3
H ₂ O ⁺	1.29	72	2.27	128	1.49	83
H ₂ O ⁻	0.16	8	0.00	—	0.40	22
CO ₂	0.27	6	0.33	7	—	—
S	0.34	—	0.13	—	—	—
SO ₃	0.85	10	Not det'd	—	—	—
P ₂ O ₅	0.09	—	0.05	—	—	—
F	0.19	11	Not det'd	—	3.70	194
Correction for Fe ₂ =0	100.30	—	—	—	101.26	—
	0.08	—	—	—	1.55	—
Total	100.22	—	99.60	—	99.71	—

¹Analyzed in the Central Chemical Lab., Uzbek Geol. Administration

²Conversion of quantitative mineral analysis to chemical.

Analytic data corroborate a general tendency in the greisenization process, as established by a microscopic study of thin sections: namely, in addition of silica and a leaching of all other components, except for alumina. The content of the latter remains approximately the same, save for the quartz zone where it is leached out, and the quartz-topaz zone where it is added.

as $\Phi = K_i + 1$, where Φ is the number of co-existing minerals, and K_i is the number of inert components.

The section is best considered going from monomineral quartz rocks to unaltered granite (Table 4).

Zone One, next to the main feeder fracture,

Table 4

Metasomatic Column

Zone No.	Mineral composition of zones		Inert components	Replacement reaction
7	Unaltered granite			Oligoclase \rightarrow albite
6	Greisenized granite	Quartz + muscovite + microcline + albite + magnetite	Al_2O_3 , K_2O Na_2O , Fe	Biotite \rightarrow muscovite + magnetite
5		Quartz + muscovite + microcline + albite	Al_2O_3 , K_2O , Na_2O	Magnetite \rightarrow leaching
4		Quartz + muscovite + microcline	Al_2O_3 , K_2O	Albite \rightarrow muscovite + quartz
3	Quartz + muscovite		Al_2O_3	Microcline \rightarrow topaz + quartz
2	Quartz + topaz		—	Microcline \rightarrow topaz + quartz
1	Quartz		All components are perfectly mobile	Topaz \rightarrow quartz

is characterized by a monomineral quartz composition. According to the phase rule, the number of the inert components (K_i) in this zone should be zero, because $\Phi = 1$. Here, we have a metasomatic zone represented by a single mineral consisting of a single perfectly mobile component, namely silica.

D. S. Korzhinskiy [5] proposes to call perfectly mobile minerals, those consisting of a single perfectly mobile component. Such a mineral is usually present in all parageneses of a metasomatic section; its amount in any zone, with a given content of inert compounds, depends on the over-all volume of the system.

Zone Two is represented by a quartz-topaz greisen. The appearance of topaz is best associated with the passage of alumina to an inert state. The extremely low mobility of alumina in metasomatic and metamorphic processes is well known and has been noted earlier by D. S. Korzhinskiy [3], V. S. Sobolev [9], V. A. Zharikov [2], and others. Chemical analyses (Table 3) show a certain enrichment of the quartz-topaz zone in alumina, as compared with unaltered granite, apparently caused by the deposition of alumina leached out of the monomineral quartz zone.

The absence of fluorite minerals in all but the quartz-topaz zones indicates its perfect mobility in greisenization processes.

Zone Three is quartz-muscovite. The presence of muscovite may be due to two causes:

1) passing of potassium to an inert state; and 2) a chemical potential of potassium, higher than in Zone Two.

As we shall see below, it is possible to determine the mobility of the third zone components by an analysis of Zone Four. In that zone, quartz, muscovite, and microcline are the stable minerals.

According to the phase rule, the number of inert components must be two. Since the alumina becomes inert as early as in Zone Two the presence of microcline can be caused only by the passiveness of potassium. Consequently muscovite, like topaz, is associated with the passivity of alumina, and its appearance instead of topaz in Zone Three is associated with an abrupt step-like rise in the chemical potential of potassium at the boundary between the second and third zones.

Farther on in the direction of the unaltered granite, albite is the stable mineral, apparently because of the passivity of sodium, while magnetite which (together with muscovite) replaces biotite is stable in segments of a very weak greisenization. The instability of biotite in all greisen rocks may be related to the high mobility of magnesium in the process of greisenization. The place of calcium in the mobility series is not quite clear; anorthite-bearing plagioclase is unstable in greisen rocks. Because of that, it has not been determined whether calcium or magnesium is the more mobile, and the latter has been tentatively

placed first in the mobility series. Considering the perfect mobility of water and carbon dioxide in all metamorphic and metasomatic processes, it can be stated that the greisenization process is realized through the following mobility series (from more to less mobile): H_2O , CO_2 , F, SiO_2 , MgO , CaO , Fe, Na_2O , K_2O , and Al_2O_3 .

The sequence of metasomatic zones and the order of mobility for their components are presented in Table 4.

SUMMARY

1. A quite definite metasomatic zonation is present in greisens associated with alaskites of the Kuramin Range.

2. A total of seven zones have been identified, from unaltered granite to monomineral quartz zones.

3. Macroscopically identifiable are three lower zones of the section: quartz, quartz-topaz, and quartz-muscovite.

The boundaries between these zones are quite distinct and the zones are identifiable right at their outcrops. The remaining zones are identifiable only in thin section, from the replacement of certain minerals by others. The boundaries between them are quite vague, with their rocks marked by a considerable amount of relict minerals. This appears to be related to the considerably lower saturation of rocks by solutions at a distance from the feeder fracture; also to the generally uneven porosity of rocks.

4. In the presence of closely spaced parallel fractures, the greisen bodies are more complex, as expressed in a multiple repetition of some zones.

5. On the basis of an analysis of metasomatic sections, and in the light of the phase rule, the following order of mobility has been established for components of a greisenization process, from more to less mobile: H_2O , CO_2 , F, SiO_2 , MgO , CaO , Fe, Na_2O , K_2O , and Al_2O_3 . The most inert component is alumina; it is mobile only in the quartz zone.

6. Present along with spatial changes in greisenizing solutions, those responsible for the zonation, are their changes in time. In connection with the decrease in acidity in solutions, leached out components (muscovite, microcline) as well as additive compounds, such as sulfides, are deposited in fractures, at terminal stages.

7. These regularities in the structure of metasomatic sections are readily explained in

the light of D. S. Korzhinskiy's theory of metasomatic processes. The absence of sharp contacts between the zones and the presence of incompletely replaced mineral relicts might have been determined by an uneven porosity of rocks, caused by fracturing in comparatively coarse-grained varieties.

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OPTIC ORIENTATION OF POTASSIUM AND SODIUM FELDSPARS AS A FUNCTION OF VARIOUS FACTORS¹

by

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Chiefly optic properties are used in identifying alkali feldspars in rocks described in geologic, petrographic, surveying, and prospecting work calling for a differentiation of igneous, metamorphic, and sedimentary sections, and in the study of feldspar mineralization associated with various ores.

These determinations are made with V. V. Nikitins's diagram of optic orientation [20] and with its rectilinear coordinate variant proposed by V. K. Monich [18]. V. V. Nikitin's diagram, however, is definitely of a preliminary nature [14]. Strictly speaking, no special analysis of optic orientation for K-Na-feldspars has been done. (Precise studies have been made of its dependence on the refractive indices in alkali feldspars [1, 23, 24, 31] and of optic orientation in comparatively rare Na-anorthoclase.)

In addition, a problem arose in connection with determining the complex phase relationships in K-Na-feldspars, on the extent to which the optic properties do reflect those relationships and the diversity of types; and that perhaps they should be replaced by X-ray methods recently worked out for these feldspars.

Sufficiently refined X-ray, electronic, and other methods have been used in recent studies of arrangement phenomena.

The feasibility of considering this factor, novel in optics, is related first of all to a higher precision in measuring the structurally-sensitive constants. We have used for that purpose the conoscopic method on Fedorov's table, with goniometrically oriented thin sections.

A correlation of these measurements with those by the standard Fedorov method on thin sections has revealed the surprising fact that the accuracy of the Fedorov method, as applied to feldspars, was exaggerated and that data so

obtained for the optic orientation of K-Na-feldspars may be considerably in error.

Because of that, we could not use the data obtained by this method (and the only data published, except for [28, 29, 36] citing some old measurements). Accordingly, we had to obtain new and sufficiently comprehensive optic data for samples analyzed both chemically and by X-ray.

In addition, there was the problem of the optics of submicroscopically twinned crystals, posed long ago by E. Mallard, and the concept of the total optics and the additivity of cryptoperthite properties.

Obtaining all these data not only enabled us to draw a new diagram of optic orientation but has led to new conclusions on the phase nature of K-Na-feldspars:

a) for K-feldspar, specifically, the optics provides means for differentiating between the effects of arrangement and sub-Roentgen twinning, not distinguishable by X-ray methods (including structural analysis);

b) in this way, new data are obtained for an interpretation of K-feldspars, supplementary to the many detailed X-ray studies; these data indicate the necessity for considering the presence of two factors determining the nature of such feldspars, their arrangement and sub-Roentgen twinning, and make it possible to guess the direction of the Al and Si displacement, in that arrangement.

I. THE STUDY METHODS

Optic orientation was determined conoscopically on the Fedorov table. Observations were made with the long-focus objective OSF 40X and a special condenser that goes with microscopes NINO4 and MIN-5. A K-feldspar fragment was first ground in a plane roughly normal to cleavage planes (001) and (010), then polished. Angles between the polished surface and planes (001) and (010) were measured on a single circle

¹Zavisimost' opticheskoy oriyentirovki kalinatriyevykh polevykh shpatov ot razlichnykh faktorov.

goniometer. The fragment was then glued, on its polished surface, to a slide glass, with the acute and obtuse angles and planes (001) and (010) marked to avoid an error, and polished down to about 0.1 mm (more or less, depending on transparency of the sample). A cover glass was not used. Such oriented sections were prepared for coarse pegmatite crystals and porphyritic granite as well as for medium-grained rocks. It was enough to break off the fragment in such a way as to have the two cleavage planes of a K-feldspar crystal at its edge.

The extremely sharp interference pattern in such thick slides was adequate for measuring the optic angle, usually down to no less than $\pm 0.5^\circ$. Axes α and γ were determined as bisectors of that angle, and β as a normal to the plane of the optic axes. An independent determination of γ was the first check of this measurement. In a preliminary estimate of angle $\perp (010) \angle \lambda$, the inclination along axis H in superposition of γ and axis I of the table was compared directly with the inclination of (010) as obtained from goniometric data. With the results plotted on the Wulfe grid, the angle between $\perp (010)$ and $\perp (001)$ was checked; it should be equal to $90^\circ \pm 0.5^\circ$, in microclines. Two series of angles were obtained as a result: 1) between axes γ , β , α , A, B, and $\perp (010)$; and 2) between these optic directions and $\perp (001)$. The projection of γ , β , α , and optic axes A and B to plane $\perp (010)$ and (001) were drawn from these data and with the Wulfe grid. This construction was an additional check of the measurement: the indicatrix axes projected must be spaced at 90° ; and optic axes A and B must lie in the $\gamma - \alpha$ plane, at the same distance from α .

Individual angles between the optic and crystallographic directions are determined in this way with high accuracy, dependent primarily on the quality of signals from (010) and (001), in goniometric measurements. The accuracy in determining the total optic orientation is somewhat lower because it depends on the error in measuring two crystallographic directions and is limited by the accuracy of graphic constructions on the Wulfe grid. In some instances, it reaches $\pm 0.5^\circ$, being generally close to $\pm 1^\circ$.

Latticed microcline cannot be measured by the conoscopic method.

Method of representation of the optic orientation. The position of optic directions in a projection to planes $\perp (010)$ and (001) is determined by the F. Becke method. A Wulfe grid is so arranged as to have its meridian arcs running from left to right. The position of a point in the projection is indicated by two coordinates, λ and φ ; $+\lambda$ is counted up from the horizontal diameter; $-\lambda$, downward from it;

$+\varphi$, to the right of the vertical diameter; and $-\varphi$, to the left. (This method was used by F. Becke for the optic orientation of plagioclase in a projection to plane $\perp (001)$.) To describe the complete optic orientation, it is sufficient to indicate the position of two optic axes A and B; the position of the indicatrix axes is readily obtainable, graphically. If angle $(010) \angle (001) = 90^\circ$, the following simple relationship will exist: $\varphi_A = 90 - \perp (010) \angle A$; $\cos (90 - \lambda_A) = \cos \perp (001) \angle A / \cos \varphi_A$. In essentially K-feldspars, angle $\alpha^* = \perp (010) \angle \perp (001)$ differs from the right angle by 25 to $30'$, in microcline. In intermediate microcline U, it is equal to $89^\circ 55'$. Consequently, in assuming $\alpha^* = 90^\circ$, for K-feldspar, we will have only a small discrepancy in the result. However, the determination of λ and φ is done more simply, graphically, with the Wulfe grid.

The extinction angle for (010), equal to $\lambda_A + (\lambda_B - \lambda_A)/2$, is readily computed for coordinates of λ .

Observation of variations in optic properties. Before measuring the optic orientation of a crystal, it is expedient to establish the presence and morphology of its variations as well as those for the optic angle, within the crystal. This is done most conveniently by observing the crystal's extinction in an oriented section. The presence of variations in the optic orientation is established first, by measuring the extinction angles, for example in section $\perp (010)$ and (001). Variations in the optic angles, not accompanied by a change in the orientation of the axes, are not present in this section. With a uniform orientation, in order to determine the presence of variations in $2V$, the slide is tilted by 20 to 30° , up or down along axis I, from (100); and by the same angle right or left along axis H. Readings of the extinction angles in that section will indicate the variation in $2V$, at a permanent orientation.

These readings of relative extinction angles are extremely sensitive to the smallest changes in optic properties within the field of vision. Being free of errors in measuring the absolute values, they readily reveal an extinction difference up to 1° . Obtained in addition is an over-all picture of the crystal from its optical properties, in addition to the approximate variation of the optic properties and their distribution throughout the crystal. Segments suitable for complete measurements should be selected on the basis of these observations.

X-ray methods. Powder patterns were obtained with the URS-50I diffractometer and Cu-radiation. The photographs were used 1) in determining the composition of homogeneous feldspars, by d_{201} (2), or in determining the degree of decay from the distance between peaks ($20\bar{1}$) of the K-feldspar and albite phases of perthite ($\Delta 2$) ($20\bar{1}$); and 2) in determining the

Table 1

Optic Orientation in K-Na-Feldspars

Sampling locality	Sample No.	$-2V^\circ$	A		B		Ng		Nm		Np		$\perp (010)$		$\perp (001)$		Extinction			
			$+\lambda$	$+\varphi$	$+\lambda$	$-\varphi$	$+\lambda$	$-\varphi$	$-\lambda$	$-\varphi$	$+\lambda$	$+\varphi$	Ng	Nm	Np	Nm	Np	$\parallel (010)$	$\perp [100] \parallel (001)$	
Sanidines																				
1	Gelkend	2330	12.0	$\parallel (010)$	11.5	0.0	0.5	0.0	0.0	90.0	5.5	0.0	90.0	0.0	90.0	5.5	90.0	84.5	5.5	0.0
2	"	2330	14.0	$\perp (010)$	5.5	7.0	5.5	7.0	0.0	90.0	0.0	84.5	0.0	5.5	0.0	0.0	90.0	90.0	5.5	0.0
2	Dva Brata	2105	6.0	$\parallel (010)$	9.0	0.0	3.0	0.0	0.0	90.0	6.0	0.0	90.0	0.0	90.0	6.0	90.0	84.0	6.0	0.0
3	Tyrny-Auz	2105	18.0		6.8	9.0	6.8	9.0	0.0	90.0	0.0	83.2	0.0	6.8	0.0	00.0	90.0	90.0	6.8	0.0
Na-anorthoclase																				
4	Dariganga I	2054	53.0		9.0	26.5	9.0	26.5	0.0	90.0	81.0	0.0	9.0	0.0	0.0	90.0	90.0	9.0	0.0	
5	Dariganga II	2053	41.0		6.8	18.0	9.5	23.0	0.0	85.8	82.0	3.5	8.0	2.3	4.2	86.5	87.7	8.2	0.0	
High orthoclase with $c \sim 2V=40-60^\circ$ (including high triclinic orthoclase)																				
6	Klyuchi	2293	39 (43)		6.6	19.5	6.6	19.5	0.0	90.0	83.4	0.0	6.6	0.0	0.0	90.0	90.0	6.6	0.0	
	"	2294	42 (43)		6.5	21.0	6.5	21.0	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	6.5	0.0	
7	Tyrny-Auz II	2322	37.5 (43)		5.5	18.75	5.5	18.75	0.0	90.0	84.5	0.0	5.5	0.0	0.0	90.0	90.0	5.5	0.0	
		2326	48.0		8.0	24.0	8.0	24.0	0.0	90.0	82.0	0.0	8.0	0.0	0.0	90.0	90.0	8.0	0.0	
		2326	52.0		6.0	28.3	9.5	23.5	60.0	86.0	82.0	3.2	8.0	2.2	4.0	86.8	87.8	7.8	0.0	
		2328	49.0		7.8	24.5	7.8	24.5	0.0	90.0	82.2	0.0	7.8	0.0	0.0	90.0	90.0	7.8	0.0	
		2328	52.0		6.0	28.0	8.0	27.3	60.0	87.8	83.0	2.0	7.0	0.7	2.2	88.0	89.3	7.0	0.0	
		2325	55.5		7.8	29.0	8.2	26.5	46.0	88.5	82.0	1.0	8.0	1.0	1.5	89.0	89.0	8.0	0.0	
		2323	53.2		6.0	27.8	10.0	25.0	70.0	86.2	82.0	3.5	8.0	1.5	3.8	86.5	88.5	8.0	0.0	
		2323	56.0		6.0	30.3	10.3	25.0	64.0	85.5	81.6	3.8	8.2	2.2	4.5	86.2	87.8	8.0	0.0	
		2323	59.0		5.0	32.5	10.0	26.3	61.0	84.5	82.0	4.5	7.8	3.2	5.5	85.8	86.8	8.1	0.0	
8	Tyrny-Auz III	2051	38.0 (41.0)		6.5	19.0	6.5	19.0	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	6.5	0.0	
9	Tyrny-Auz IV	2027	43.0		6.5	21.5	6.5	21.5	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	6.5	0.0	
		2027	56.0		8.5	28.0	8.5	28.0	0.0	90.0	81.5	0.0	8.5	0.0	0.0	90.0	90.0	8.5	0.0	

Table 1 continued

Sampling locality	Sample No.	-2V°	A		B		Ng		Nm		Np		∠ (010)			∠ (001)			Extinction			
			+λ	+φ	+λ	-φ	+λ	-φ	-λ	-φ	+λ	+φ	Ng	Nm	Np	Ng	Nm	Np	∥ (010)	∠ [100]	∥ (001)	
			Intermediate microcline and orthoclase, -2V=60-80°																			
10	Ortotokey I	2156	49.0	24.5	6.0	24.5	0.0	90.0	84.0	0.0	6.0	0.0	0.0	90.0	90.0	90.0	6.0	84.0	6.0	0.0	0.0	
		2154	52.5	26.3	6.5	26.3	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	90.0	6.5	83.5	6.5	0.0	0.0	
11	Ortotokey II	2230	58.0	29.0	5.5	29.0	0.0	90.0	84.5	0.0	5.5	0.0	0.0	90.0	90.0	90.0	5.5	84.5	5.5	0.0	0.0	
12	Kukisvumchorr I	2179	55.0	27.5	6.0	27.5	0.0	90.0	84.0	0.0	6.0	0.0	0.0	90.0	90.0	90.0	6.0	84.0	6.0	0.0	0.0	
		2173	56.0	27.0	5.8	27.0	30.0	87.7	84.8	1.0	5.3	2.0	2.3	89.0	88.0	88.5	5.3	85.0	5.2	1.0	1.0	
		2166	58.0	26.8	6.5	26.8	48.0	87.0	84.3	2.0	5.5	2.2	3.0	88.0	87.8	87.9	5.8	84.8	5.3	3.0	2.0	
13	Cherepashino	2246	61.0	33.5	5.5	27.0	38.0	86.0	83.0	2.0	6.8	3.0	4.0	88.0	87.0	88.0	7.0	83.3	7.4	3.3	3.50	
14	Il'meny	2287	62.0	31.0	6.3	31.0	0.0	90.0	83.7	0.0	6.3	0.0	0.0	90.0	90.0	90.0	6.3	83.7	6.3	0.0	0.0	
15	Ankavan	2318	63.0	32.8	8.0	30.0	60.0	88.0	80.0	1.5	10.0	1.5	2.0	88.5	88.5	88.5	9.8	80.2	9.6	2.0	1.5	
		2142	62.0	31.0	7.0	31.0	0.0	90.0	83.0	0.0	7.0	0.0	0.0	90.0	90.0	90.0	7.0	83.0	7.0	0.0	0.0	
16	Amovka	2192	58.0	28.0	6.2	28.0	58.0	84.5	1.8	5.5	1.0	2.0	2.0	88.2	89.0	88.0	5.5	84.8	5.1	2.0	1.0	
		2192'	65.0	3.0	35.6	7.0	29.6	45.0	85.0	3.5	5.7	3.5	5.0	86.5	86.5	86.8	6.8	85.0	5.0	3.8	3.0	
		2021	62.5	4.2	34.5	7.2	28.0	32.0	85.8	2.0	6.0	3.8	4.2	88.0	86.2	88.0	6.0	84.0	5.7	4.2	3.5	
		2021	69.0	4.0	41.0	7.6	28.3	30.0	83.5	84.0	1.5	6.0	6.5	87.5	84.0	86.8	6.2	84.5	5.8	4.0	6.0	
		2021	73.5	4.8	44.0	9.2	29.0	38.0	81.0	4.8	6.0	7.0	9.0	85.2	82.5	84.5	7.5	85.0	5.5	7.0	7.5	
		2128	65.5	3.0	37.0	9.7	28.0	55.4	82.7	5.7	6.2	4.5	7.3	84.3	85.5	84.0	8.6	83.8	6.3	6.0	4.5	
17	Kukisvumchorr II	2151	62.0	5.8	33.0	6.8	29.0	40.0	88.7	3.2	1.5	6.2	1.5	2.0	88.5	88.5	88.5	6.5	83.7	6.3	2.0	4.0
		2217	63.5	4.0	35.0	8.0	28.5	47.0	85.8	84.0	3.0	3.0	4.2	87.0	87.0	86.8	6.5	84.3	6.0	4.2	2.0	
		2217	63.0	4.4	32.5	6.4	30.5	50.0	88.0	84.7	1.7	5.3	1.0	2.0	88.3	89.0	88.5	5.3	85.0	5.4	2.0	1.8
18	Ortotokey III	2157	66.0	5.5	34.0	6.0	32.0	48.0	88.2	1.5	5.7	1.0	1.8	88.5	89.0	89.0	5.6	84.5	5.9	1.5	1.2	
19	Taymyr I	2132	62.0	6.5	34.0	6.5	31.0	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	90.0	6.5	83.5	6.5	0.0	0.0
	(syenite)	2131	66.0	7.3	33.0	7.3	33.0	0.0	90.0	82.7	0.0	7.3	0.0	0.0	90.0	90.0	90.0	7.3	82.7	7.3	0.0	0.0
20	Buzheninov Bor	2305	69.5	6.0	40.5	12.0	29.0	49.5	82.2	80.0	5.0	9.8	6.0	7.8	85.0	84.0	84.3	11.0	80.5	9.0	7.0	5.2
		2026	66.0	5.0	35.2	8.2	30.5	50.0	86.5	83.3	2.5	6.8	2.5	3.5	87.5	87.5	87.5	7.0	83.5	6.6	3.0	3.0

Table 1 continued

	Sampling locality	Sample No.	-2v°	A		B		Ng		Nm		Np		⊥ (010)			⊥ (001)			Extinction		
				+λ	+φ	+λ	-φ	+λ	-φ	-λ	-φ	+λ	+φ	Ng	Nm	Np	Ng	Nm	Np	[010]	⊥ [100]	[001]
21	Lovozero	2227	67.0	6.0	33.5	6.0	33.5	0.0	90.0	84.0	0.0	6.0	0.0	0.0	90.0	90.0	90.0	6.0	84.0	6.0	0.0	
		2298	68.0	10.0	35.0	12.0	32.7	64.0	88.0	79.0	1.5	10.8	1.5	2.0	88.5	88.5	88.5	10.7	79.5	11.0	2.0	
		2297	78.0	8.0	40.7	13.3	37.0	70.0	86.0	79.8	3.5	10.5	2.0	4.0	86.5	88.0	86.5	10.8	79.8	10.6	4.0	
		2234	65.5	7.5	32.75	7.5	32.75	0.0	90.0	82.5	0.0	7.5	0.0	0.0	90.0	90.0	90.0	7.5	82.5	7.5	0.0	
23	Gnivan	2082	75.5	6.5	37.75	6.5	37.75	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	90.0	6.5	83.5	6.5	0.0	
		2136	66.5	4.5	39.0	9.5	27.3	35.0	83.3	82.3	3.5	7.0	5.8	6.7	86.5	84.2	86.0	8.2	82.7	7.0	5.5	
		2163	74.5	7.0	37.25	7.0	37.25	0.0	90.0	83.0	0.0	7.0	0.0	0.0	90.0	90.0	90.0	7.0	83.0	7.0	0.0	
		2055	67.0	3.8	36.3	9.0	30.3	58.0	85.0	84.0	4.0	6.3	3.0	5.0	86.0	87.0	86.0	7.2	84.0	7.6	4.8	
24	Zhezhelev	2135	66.5	6.0	34.2	7.8	32.2	50.0	88.2	83.0	1.5	6.8	1.0	1.8	88.5	89.0	89.0	6.8	83.3	6.9	1.7	
		2246	72.0	7.5	36.0	7.5	36.0	0.0	90.0	82.5	0.0	7.5	0.0	0.0	90.0	90.0	90.0	7.5	82.5	7.5	0.0	
		2301	72.5	6.3	37.8	9.0	34.2	50.0	87.5	82.0	2.0	8.0	1.5	2.5	88.0	88.5	88.0	8.0	82.2	7.6	2.3	
		2247	73.0	7.5	36.5	7.5	36.5	0.0	90.0	82.5	0.0	7.5	0.0	0.0	90.0	90.0	90.0	7.5	82.5	7.5	0.0	
25	Strizhevka	2248	73.0	4.0	38.5	8.5	34.7	60.0	86.5	83.7	3.0	6.5	2.0	3.5	87.0	88.0	87.0	7.5	82.5	7.6	2.0	
		2165	68.0	7.5	34.0	7.5	34.0	0.0	90.0	82.5	0.0	7.5	0.0	0.0	90.0	90.0	90.0	7.5	82.5	7.5	0.0	
		2245	68.0	5.0	37.0	8.5	31.0	41.0	86.0	83.0	2.5	7.0	3.0	4.0	87.5	87.0	87.3	7.3	83.3	6.7	3.0	
		2058	70.0	4.0	41.7	10.0	28.0	38.0	82.0	82.0	4.0	7.8	7.0	8.0	86.0	83.0	85.5	8.2	82.3	7.6	4.0	
26	Sabarov	2244	73.0	2.0	44.5	11.0	28.0	42.0	79.8	82.0	6.0	7.0	8.2	10.2	84.0	81.8	83.2	9.7	83.2	6.5	8.8	
		2036	72.0	4.0	39.3	10.0	32.5	52.0	85.0	82.5	4.0	7.5	3.0	5.0	86.0	84.5	85.5	8.0	83.0	7.0	4.8	
		2311	58.0	4.0	34.0	8.0	23.5	41.0	83.2	83.5	4.0	6.0	5.5	6.8	86.0	84.5	85.5	7.5	84.0	6.0	5.0	
		2309	62.5	3.0	38.0	9.5	24.0	41.0	81.5	82.6	5.0	6.5	7.0	8.5	85.0	83.0	84.5	8.8	83.2	6.2	8.0	
27	Taymyr II (pegmatite)	2309	66.0	2.0	40.0	10.0	25.0	44.0	80.0	82.5	6.3	6.5	7.7	10.0	83.7	82.3	82.8	9.8	83.3	6.0	9.0	
		2308	59.0	3.3	35.0	9.0	23.0	42.0	82.5	83.0	4.8	6.3	5.8	7.5	85.2	84.2	84.8	8.0	84.0	6.0	7.0	
		2308	61.0	2.5	37.0	9.0	24.0	43.0	81.8	83.2	5.2	6.0	6.3	8.2	84.8	83.7	84.2	8.3	84.0	5.7	7.3	
		2308	65.5	1.7	43.0	10.0	24.5	41.0	80.0	83.0	6.0	6.2	8.0	10.0	84.0	82.0	83.3	9.0	84.0	5.8	9.5	
2308	74.5	-2.0	49.0	12.0	25.0	40.0	75.5	82.0	8.0	6.0	12.0	14.5	82.0	78.0	80.8	11.0	84.0	5.0	12.5			

Table I continued

Sampling locality	Sample No.	-2V°	A		B		Ng		Nm		Np		⊥ (010)			⊥ (001)			Extinction				
			+λ	+φ	+λ	-φ	+λ	-φ	-λ	-φ	-λ	+φ	Ng	Nm	Np	Ng	Nm	Np	(010)	⊥ {100}	(001)		
28	Malaya Viska	2310	-0.5	45.0	10.0	24.5	40.0	77.7	83.2	7.0	5.5	10.0	12.3	83.0	80.0	82.0	9.8	84.5	5.2	10.5	40.0		
		2309	-4.0	57.0	42.7	25.0	35.0	71.8	81.2	8.5	6.0	16.0	18.2	81.5	74.0	79.6	12.0	84.0	4.3	14.0	16.0		
		2266	70.5	8.0	36.0	10.5	35.0	60.0	87.5	80.5	2.0	9.3	1.5	2.5	88.0	88.5	88.0	9.5	80.8	9.1	2.0	1.0	
		2267	74.0	9.5	37.0	9.5	37.0	0.0	90.0	80.5	0.0	9.5	0.0	0.0	90.0	90.0	90.0	9.5	80.5	9.5	0.0	0.0	
		2221	70.5	7.8	36.5	8.2	33.5	30.0	88.3	82.0	0.5	8.0	1.5	1.7	89.5	88.5	89.3	8.1	82.0	8.0	1.0	1.5	
		2262	73.5	7.0	39.0	10.0	35.5	56.0	87.2	81.5	2.0	8.5	2.0	2.8	88.0	88.0	88.0	8.7	81.5	8.5	2.0	1.2	
		31 Shpola	2264	77.5	5.0	44.7	10.0	32.7	34.0	83.5	82.0	3.3	7.7	5.7	6.5	86.7	84.3	86.3	8.3	82.5	7.5	5.0	6.0
		32 Korsun'	2255	80.0	4.0	43.5	11.0	36.0	55.0	84.5	82.0	4.0	8.0	4.0	5.5	86.0	86.0	85.5	9.0	82.3	7.5	5.5	3.0
		2256	80.5	8.3	42.5	10.2	38.0	35.0	88.0	80.2	1.2	9.8	2.0	2.0	88.8	88.0	89.0	9.8	80.2	9.3	1.5	1.8	
		33 Malin	2111	80.0	7.0	42.2	12.2	37.5	59.0	86.0	80.0	3.3	10.0	2.5	4.0	86.7	87.7	86.5	10.3	80.3	9.6	3.8	2.5
34 Kapustino	2193	72.0	6.2	42.0	11.0	30.0	83.3	83.3	80.3	1.7	9.2	6.0	6.7	86.8	84.0	86.0	9.8	81.0	8.6	5.8	6.0		
	2194	80.0	9.0	43.3	11.8	36.8	39.0	86.0	80.0	1.8	10.0	3.3	4.0	88.2	86.7	88.5	10.3	80.0	10.4	2.8	3.3		
Cryptocrystalline microclines																							
35	Voinovka	2017	82.0	8.0	44.0	11.5	37.5	86.0	80.0	2.0	10.0	3.5	4.0	88.0	86.5	87.5	10.0	80.2	9.7	3.3	3.0		
		2017	83.0	5.0	50.0	12.8	32.8	35.0	80.5	80.0	4.0	9.5	8.5	9.5	86.0	81.5	85.0	10.5	80.8	8.9	7.5	8.5	
		2191	82.5	8.0	43.5	8.4	39.0	29.0	87.9	81.9	0.5	8.1	2.0	2.1	89.5	88.0	89.3	8.1	82.0	9.0	3.0	2.0	
		2188	84.0	7.5	46.0	10.5	38.0	38.0	85.7	80.4	1.5	9.6	4.0	4.3	88.5	86.0	88.0	9.5	80.6	9.0	3.0	4.0	
		2189	84.5	5.0	48.3	11.0	36.0	33.0	83.2	81.0	3.0	9.0	6.0	6.8	87.0	84.0	86.2	9.5	80.5	8.0	5.0	6.2	
		2189	84.0	2.5	49.5	11.5	34.0	48.0	80.5	80.8	4.6	8.0	8.0	9.5	85.5	82.0	84.5	9.5	82.2	7.0	6.8	8.0	
		2190	84.5	8.4	44.5	11.0	40.0	30.0	87.5	80.1	1.5	10.0	2.5	2.5	88.5	87.5	88.2	10.0	80.0	9.7	2.0	2.5	
		2204	83.0	3.8	49.8	12.0	33.0	39.3	80.5	81.3	5.0	8.1	8.0	9.5	85.0	82.0	84.0	10.0	82.0	7.9	7.3	8.5	
		2205	83.0	4.0	52.0	11.8	31.5	20.0	79.0	80.5	4.0	8.2	10.0	11.0	86.0	86.0	85.2	10.1	82.0	7.9	7.7	10.0	
		2205	84.5	6.0	43.7	8.0	39.5	29.0	87.8	83.0	1.0	7.0	2.0	2.2	89.0	88.0	89.2	7.0	83.0	7.0	1.3	2.0	
36	Adabash	2205	84.5	2.0	53.0	12.0	31.5	29.0	81.0	5.0	8.0	10.8	12.0	85.0	79.1	83.8	10.0	82.2	7.0	8.2	10.5		

Table 1 continued

	Sampling locality	Sample No.	-2V°	A		B		Ng		Nm		Np		⊥ (010)		⊥ (011)		Extinction			
				+λ	+φ	+λ	-φ	+λ	-φ	-λ	-φ	+λ	+φ	Nλ	Nm	Np	Nλ	Nm	Np	(010)	⊥ [100]
37	Kirovgrad	2268	84.0	3.0	51.5	11.0	32.5	32.0	79.5	81.8	4.7	8.0	9.5	10.5	85.5	80.5	84.5	9.5	82.3	7.0	9.5
		2268	85.0	4.5	51.0	11.8	33.5	30.0	80.0	80.5	4.0	9.0	9.0	10.0	86.0	81.5	85.0	10.0	81.5	8.0	6.8
		2269	80.5	4.0	45.5	9.8	39.0	45.0	85.7	83.0	2.8	7.0	3.0	4.3	87.0	87.0	87.0	7.3	83.5	7.9	3.5
		2063	80.0	8.0	41.0	10.0	39.0	55.0	88.0	80.3	1.5	9.5	1.0	2.0	88.3	89.0	89.5	9.3	80.8	9.0	1.8
		2064	82.0	8.0	42.0	8.8	40.0	60.0	88.0	82.3	1.5	7.5	1.5	2.0	88.5	88.5	88.0	7.5	82.8	7.9	2.0
39	Migiya	2275	81.0	4.0	42.0	13.0	38.5	60.0	83.0	5.5	8.7	2.0	7.0	84.5	88.0	84.2	10.2	81.8	8.5	5.3	
Adularia																					
40	Astaŭyevskoye	2162	36.0	5.5	18.0	5.5	18.0	0.0	90.0	84.5	0.0	5.5	0.0	0.0	90.0	90.0	90.0	5.5	84.5	5.5	0.0
		2162	40.0	5.5	20.0	5.5	20.0	0.0	90.0	84.5	0.0	5.5	0.0	0.0	90.0	90.0	90.0	5.5	84.5	5.5	0.0
		2220	62.0	4.0	34.0	6.3	28.0	38.0	86.0	84.5	2.0	5.5	3.5	4.0	88.0	86.5	87.8	5.6	84.8	5.2	3.0
		2100	63.0	2.5	38.2	8.0	24.2	36.0	82.0	84.0	4.0	5.5	7.0	8.0	86.0	83.0	85.3	7.0	84.8	5.2	6.2
		Bututychag	2100	44.5	5.5	22.25	5.5	22.25	0.0	90.0	84.5	0.0	5.5	0.0	0.0	90.0	90.0	90.0	5.5	84.5	5.5
41	Bututychag	2100	53.0	5.5	26.5	5.5	26.5	0.0	90.0	84.5	0.0	5.5	0.0	0.0	90.0	90.0	90.0	5.5	84.5	5.5	0.0
		2099	53.5	6.5	26.75	6.5	26.75	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	90.0	6.5	83.5	6.5	0.0
		2099	58.5	6.5	29.25	6.5	29.25	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	90.0	6.5	83.5	6.5	0.0
		2099	62.0	6.5	31.0	6.5	31.0	0.0	90.0	83.5	0.0	6.5	0.0	0.0	90.0	90.0	90.0	6.5	83.5	6.5	0.0
		Barium orthoclase																			
42	Inaglin	2092	57.0	1.7	28.5	1.7	28.5	0.0	90.0	88.3	0.0	1.7	0.0	0.0	90.0	90.0	90.0	1.7	88.3	1.7	0.0
		2130	57.0	1.2	28.75	1.2	28.75	0.0	90.0	88.8	0.0	1.2	0.0	0.0	90.0	90.0	90.0	1.2	88.8	1.2	0.0
		2306	73.5	1.0	36.75	1.0	36.75	0.0	90.0	89.0	0.0	1.0	0.0	0.0	90.0	90.0	90.0	1.0	89.0	1.0	0.0
43	Slyudyanka	Nonlattice microcline																			
44	Rischorr	2184	83.5	-5.0	56.5	13.0	26.5	36.5	72.0	81.5	9.0	6.9	15.5	18.0	81.0	74.5	79.0	12.0	84.2	4.0	14.0
		2182	84.5	-5.0	57.2	13.0	25.7	36.0	71.8	81.0	9.0	6.3	15.8	18.2	81.0	74.2	79.5	12.5	83.2	4.0	14.5
45	Zaporozh'ye																				

Table 2
Chemical Composition and X-ray Data for the Study of Samples

Sample No.	Sampling location	Chemical composition					X-ray diffractometer				X-ray variations	
		Or	Ab	An	Cs (Ba)	$\Delta 2\theta$ (2 θ T)	d	Δ	K-phase	α	γ	Na-phase
1	Gelkend	76.7	23.3	—	—	—	4.1866	0.0	—	—	—	—
3	Tyrny-Auz I	72.4	27.9	—	(0.2)	—	4.1788	0.0	—	—	—	—
4	Dariganga	26.5	68.4	3.3	—	—	—	—	—	—	—	—
6	Klyuchi	—	—	—	0.4	1.08	—	0.0	—	—	86°00'	88°50' Ac
7	Tyrny-Auz II	66.2	30.8	2.6	(0.00n)	1.08	—	0.0	—	—	—	—
8	Tyrny-Auz III	67.3	31.4	1.3	—	—	—	0.0	—	—	—	—
9	Tyrny-Auz IV	82.5	17.5	—	—	—	4.2181	0.0	—	—	86°05 A	88 58 A—Pc —Pc
10	Ortotokoy I	75.4	24.6	—	—	1.08	—	0.0 diff.	1	diff	—	—
11	Ortotokoy II	66.2	33.8	—	—	—	—	—	1	diff	A	—
12	Kukisvumchorr	188.2	11.8	—	(0.0n)	—	4.2221	0.0	—	—	—	—
13	Cherepashino	—	—	—	—	—	—	0.0	—	—	—	—
14	Il'meny	80.5	19.5	0.5	0.2	1.13	—	0.0 diff.	1	diff	—	—
15	Ankavan	—	—	—	—	1.10	—	0.0	—	—	86°21'	89°08' A-Pc
16	Annovka	88.2	11.8	—	—	1.13	—	0.0 diff.	1	diff	A	—
17	Kukisvumchorr II	82.7	17.05	0.05	0.2	—	4.2221	0.0	—	—	—	—
18	Ortotokoy III	—	—	—	—	1.10	—	0.0	—	—	86°11'	90°45' A-Pc
20	Buzheninov Bor	68.7	30.6	0.5	0.2	1.14	—	0.0	—	—	86°08'	90°39' A
21	Lovozero	75.5	20.8	3.4	0.3	1.09	—	0.0—0.70	1	diff	—	—
22	Tayezhnoye	73.2	26.2	0.4	0.2	1.14	—	0.0	—	—	—	—
23	Gnivan'	71.5	25.5	3.0	(0.2)	1.12	—	0.0	—	—	—	—
24	Zhezhelev	72.4	25.5	1.3	0.8	1.10	—	0.0—0.36	1	diff	86°38'	90°25' A
25	Strizhevka	75.3	21.5	3.2	(0.2)	—	—	—	1	diff	A	—
26	Sabarov	77.5	21.0	1.5	(0.2)	—	—	—	1	diff	—	—
27	Taymyr II	68.9	31.0	0.1	(0.00n)	1.09	—	0.0 diff.	1	diff	—	—
28	Malaya Viska	—	—	—	—	1.10	—	0.0 diff.	1	diff	A	—
30	Khlystunovka	64.3	35.2	—	—	1.14	—	0.0 diff.	—	—	86°23'	90°40' A-P
31	Shpola	64.1	34.7	—	(0.2)	1.14	—	0.0 diff.	1	diff	86°31'	90°36' A-P
32	Korsun'	—	—	—	—	1.12	—	0.0—0.80	1	diff	86°42'	90°30' A-P
34	Kapustino	—	—	—	—	1.11	—	0.0—0.72	1	diff	86°20'	90°23' A
35	Voinovka	65.0	34.0	1.0	(0.2)	1.10	—	0.0—0.99	1	diff	86°10'	90°31' A-P
36	Adabash	73.1	25.3	1.6	(0.2)	1.14	—	0.98	1	diff	86°20'	90°41' A
37	Kirovgrad	66.5	30.5	3.0	—	1.09	—	0.90	1	diff	—	—
39	Migiya	71.1	28.5	0.4	(0.2)	—	—	—	—	—	—	—
40	Astaf'yevskoye	—	—	—	—	—	4.2261	0.0	—	—	—	—

Sample No.	Sampling location	Chemical composition					X-ray diffractometer				X-ray variations	
		Or	Ab	An	Cs (Ba)	$\Delta 2\theta$ (201)	d	Δ	K-phase α^*	Na-phase γ^*	α^*	γ^*
41	Bututychag	91.2	8.8	—	—	—	—	—	—	—	—	—
42	Inaglin	82.3	11.8	1.2	4.4	—	—	0.0	—	—	—	—
43	Slyudyanka	70.6	17.9	1.1	10.4	1.12	—	0.0	1	—	A(?)	—
44	Rischorr	87.95	12.0	0.05	(0.n)	1.10	—	0.99	1	—	—	—
45	Zaporozh'ye	—	—	—	—	—	—	—	—	—	—	—
46	Karelia	78.5	19.4	2.1	—	1.09	—	0.98	90°27' 92°18' A	—	—	—
47	Azov region	—	—	—	—	1.10	—	0.99	—	—	—	—

A - albite twinning; P - periclinal twinning; c - twinning superstructure; I - irrational twinning; 1 - one series of K-phase reflections. Angles α^* and γ^* measured by S.V. Rykova. Column Cs (Ba), in parentheses, gives Ba content, from spectrographic analysis.

degree of triclinity [5] and the crystal symmetry by reflection (131) and also (130).

X-ray photographs of variations about axis b^* , with plane (001) located in the middle of the 15° variation angle [38], were used in determining the homogeneous or submicroscopic perthite structure of the sample and the presence and nature of twinning laws for both phases of this perthite; in the presence of distinct spots, angles α^* and γ^* were measured in twinning positions. Such photographs were obtained by S. V. Rykova. She measured the same angles in the non-twinning sample "Taymyr-II", by the Weissenberg method of evolutes of zero layer lines.

All photographs made by the monocrystal methods were obtained for fragments of about 0.1 mm, taken out of an oriented section, after the optic orientation had been measured at that very spot.

Chemical analyses. Identification of K_2O , Na_2O , and CaO was done by the flame photometric method (Central Chemical Laboratory, I. G. E. M.).

A thorough chemical analysis was done on four samples; five samples, accompanied by their chemical data, were obtained from G. A. Afanas'yev, V.I. Vlodavets, T.I. Kozhina, S. M. Kravchenko, and Ye.I. Semenova; and one analysis was borrowed from the literature (S. P. Solov'yev).

II. THE RESULTS OF STUDY

Samples from 130 different localities were studied in more or less detail, with 47 selected for further study.

The results of optic, chemical and x-ray analyses are presented in Tables 1 and 2. (Data on optic orientation of latticed microclines are not cited because they cannot be used for control.) Material is arranged in ascending order of optic angles.

A description of these samples and of their geologic position is given in our other work [16].

III. ANALYSIS OF THE RESULTS

Inaccuracy of the Orthoscopic Method of Measuring the Optic Orientation of K-Na-Feldspars

The errors in measuring K-feldspars in thin section by the standard Fedorov method arise from a dual cause.

First, there is the error in determining the crystallographic directions from cleavage. In plagioclases, the accuracy in determining the inclination of a cleavage plane reaches 2 to $3^\circ \pm$. In K-feldspars, whose refractive index is closer to that of Canada balsam, this accuracy is reduced for the (001) cleavage, and even more so for the weaker (010) cleavage, often missing in sections.

Second, there is the error in determining the position of the optic indicatrix. Because of low birefringence and a considerable dispersion, determinations by the extinction method are also less accurate for K-Na-feldspar than for plagioclase.

However, the error in determining the indicatrix elements rises abruptly in measuring K-feldspars with moderate to small optic angles. With birefringence $\beta - \alpha = 0.005$ and $2V = -50^\circ$, the difference between γ and β is but 0.001 , and 0.0003 for $2V = -30^\circ$. Sanidines with $2V \approx 20^\circ$ behave like uniaxial minerals on the Fedorov table, so that any attempt at an orthoscopic determination of γ and β in them is the same as trying to find these axes in an uniaxial mineral.

The question arises — how large is this error in measurement? It is not constant, being dependent on the excellence of cleavage, the thickness of the slide, the selection of the cross section, and particularly on the magnitude of the optic angle (the possible error is greatest for small $2V$'s).

The error can be estimated by comparing the standard measurements with the conoscopic, on the Fedorov table, in goniometrically oriented preparations.

For an illustration, we shall take the measurements on K-feldspars (Figure 1) from the Tyrny-Auz porphyritic granite [4]. This is one of the K-feldspars measured in the most detail and its data are used as typical [19]. The author's authority and experience [4] preclude any errors save those inherent in the method itself. The most conspicuous is the dispersion of points (Figure 1) obtained by the standard Fedorov method. Without the assumption of an incidental nature of optic constants in K-feldspars, not related to composition and structure, such a dispersion of points would mean that the rock contains K-feldspars of the most diverse types, quite different composition, and with almost extreme structural situations. A comparison of points corresponding to the highest microcline, arranged and non-arranged albite, and submicroscopically twinned varieties, shows that individual variations correspond to a content of more than 100% albite component with a higher triclinitly than the highest microcline. A comparison of these data and conoscopic measurements in

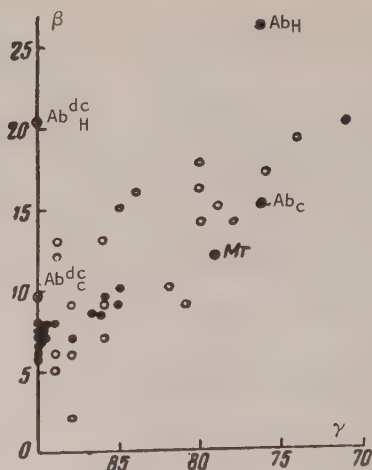


FIGURE 1. K-feldspar from the Eldzhurtin granite (Tyrny-Auz II, No. 3, Table 1).

Blank circles — orthoscopic measurements [5]; 23 measurements in 6 sections. Blank circles — conoscopic measurements on the Fedorov table, in goniometrically oriented tablets (author's data); 21 measurements on 17 crystals. Crossed circles — points of microcline (M_r), "low-" and "high-temperature" albite (Ab_C , Ab_D), and their submicroscopically twinned crystals (Ab_H , Ab_E).

goniometrically oriented sections of samples from the same massif shows that the error in measuring individual angles by the standard method reaches 10° and more.

A similar dispersion of values and the presence of optic orientation data impossible for K-Na-feldspars have also been observed in other works published during the last 40 years, as the results of many thousands of measurements. As the inaccurate measurements accumulated, they gave birth to certain concepts which became psychologic factors in propagating these erroneous data. This is especially true for works where K-feldspars were correlated by a comparison of maxima obtained through a statistical analysis of inaccurate measurements. Such statistics are mostly the statistics of errors.

One hears sometimes that a statistical analysis of a large number of inaccurate measurements may come up with a correct average result. However, in our study, all deviations were found to be in the same direction. For example, inaccurate measurements of samples with a monoclinic orientation, and resulting in a deviational triclinic orientation, will never produce an average monoclinic value but rather an average deviation from it.

Total Optics, Additivity of Properties, and Determination of the Composition of Microperthite

The observed optic properties of cryptoperthite feldspars are composite, representing the combined properties of a twinned K-phase and a Na-phase also twinned as a block crystal; each of these phases may have a different degree of arrangement. Thus the optic properties of cryptoperthites represent a complex function of several factors: 1) over-all composition of feldspar; 2) its homogeneous or perthitic state; 3) the composition of each phase; 4) submicroscopic twinning; and 5) the degree of arrangement in each phase.

Extinction, conoscopic figure, signals in measuring the refractive indices by the prism method, the Becke band, are all results of the total optic effect in complexly built block crystals.

Perthites of a substantially advanced arrangement consist [16] of phases similar to pure components. For this reason, the optic properties of cryptoperthites may be assumed, in the first approximation, to depend solely on the quantitative ratio of perthite phases, by taking the phase composition to be similar to terminal members.

As far as physicochemical analysis is concerned, the relationship between the composition and properties is different for homogeneous and decayed solid solutions. In the first, it is expressed generally by a more or less complex curve. In the second, it is expressed by a straight line (of the mechanical mixture type).

However, the additivity of properties is not an individual feature but a characteristic of pseudo-monoclinic twinned block crystals of K- and Na-phases with different structural setups. We have computed the optic properties of such block crystals (see below).

These considerations were used in constructing a diagram for determining from total optic properties, the over-all composition of cryptoperthite feldspars similar to pseudomonoclinic ([15], Figure 2).

The Effect of Submicroscopic and Cryptotwinning on Optic Properties of K-Na-Feldspars

According to the elementary hypothesis promulgated by E. Mallard [33] in 1876, "orthoclase is cryptomicrocline" (in D. S. Belyankin's formulation, [2]). Direct demonstrations of the existence of optically homogeneous although submicroscopically twinned alkali feldspars have been obtained by X-ray methods [9, 10, 16, 32, 37].

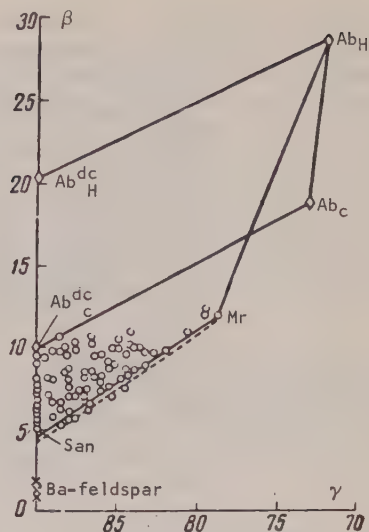


FIGURE 2. The position of 1 (001) relative to γ , β , and α of the K-Na-feldspar studied

An attempt to calculate the optic properties of "orthoclase" which was a submicroscopically twinned microcline was made by T. F. Barth [26]. According to him, "orthoclase" with $2V = -67.5^\circ$ was obtained from microcline with $2V = -75^\circ$, through submicroscopic twinning. However, as noted by H. Oftung [21] that calculation was done with Mallard's formula, inapplicable to pseudo-monoclinic specimens; the author of the submicroscopic twinning hypothesis, himself, did not apply his formula in computing the orthoclase-microcline ratios.

We have described elsewhere [12] a general method of computing the optic properties of submicroscopically twinned balanced block crystals, from individual optic properties.

Table 3 presents the optic properties for certain alkali feldspars, and the properties of submicroscopically twinned block crystals whose components they are computed from.

A petrographically important result of this calculation is that "common orthoclase" with $-2V = 60$ to 70° is not a submicroscopically twinned maximum microcline with $-2V$ of about 84° . There are parallel series of K-feldspars differing in symmetry, triclinic in some, pseudo-monoclinic in others, with almost the same optic angles: in pseudo-monoclinic block-crystals, $2V$ is reduced by about 3° compared with individual crystals. "Orthoclase" consists of submicroscopically twinned block crystals consisting of individuals with various arrangements. We have constructed a graph showing the relationships of the

Table 3
Results of Computation of Optic Properties of Certain
Submicroscopically Twinned Alkali Feldspars

Sample No.	Mineral	Ng	Np	2V	A		B		Ng		Nm		Np		⊥ (010)		⊥ (001)	
					λ	φ	λ	φ	λ	φ	λ	φ	λ	φ	Ng	Nm	Np	Np
1	Microcline (36)	1.5260	1.5200	-84.0	-4.0	+57.0	+12.8	-25.5	+35.5	-72.0	-81.0	-8.0	+7.0	+16.0	18.0	82.0	74.0	80.0
1a	Twinned microcline	1.52542	1.52050	-81.0	+4.5	+40.5	+4.5	-40.5	0.0	90.0	-85.5	0.0	+4.5	0.0	0.0	90.0	90.0	90.0
2	Intermediate microcline	1.5255	1.5195	-74.0	0.0	+49.3	+10.3	-24.0	+33.5	-75.8	-82.3	-6.3	+6.2	+12.7	14.2	83.7	77.3	82.3
2a	Twinned intermediate microcline	1.52516	1.51967	-70.5	+5.1	+35.2	+5.1	-35.2	0.0	90.0	-84.9	0.0	+5.1	0.0	0.0	90.0	90.0	90.0
3	Intermediate microcline	1.5256	1.5203	-66.0	+3.0	+43.0	+9.5	-22.5	+31.0	-79.0	-82.5	-4.7	+6.3	+10.0	11.0	85.3	80.0	84.3
3a	Twinned intermediate microcline	1.52542	1.52046	-63.5	+6.2	+31.7	+6.2	-31.7	0.0	90.0	-83.8	0.0	+6.2	0.0	0.0	90.0	90.0	90.0
4	Albite	1.5392	1.5291	+78.7	0.0	+48.5	+39.0	-47.2	-72.2	+73.0	-70.0	-17.0	-20.0	+0.2	17.0	73.0	89.8	73.8
4a	Twinned albite	1.53868	1.52911	+86.8	+20.5	+43.4	+20.5	-43.5	0.0	90.0	-79.5	0.0	+20.5	0.0	0.0	90.0	90.0	90.0
5	Analcite (37)	1.5341	1.5272	-50.0	4.0	+12.3	+20.5	-32.0	-64.0	+61.5	-88.0	-26.7	+7.3	-10.0	28.5	63.3	80.0	64.2
5a	Twinned analbite	1.53360	1.52740	-32.0	+9.7	+16.0	+9.7	-16.0	0.0	90.0	-80.3	0.0	+9.7	0.0	0.0	90.0	90.0	90.0

K-feldspars ([15], Figure 1). In albite, especially in "high-temperature" albite, the difference in optic angles between individuals and twinned block crystals increases to 10 or 20° (see Table 3).

Structural Significance of Optic Properties in K-Na-Feldspars

All gradual changes in properties, with a constant composition, are related to the arrangement phenomena. The effect of arrangement on electric, magnetic, and mechanical properties is treated in the voluminous physical and metallographical literature [7].

The dependence of properties described by Fresnel's crystal optics, on arrangement has been considered in the light of physicochemical analysis [11].

Feldspars present an outstanding example of the effect of arrangement on optic properties. In considering the optic orientation of plagioclases [13] we used the extreme values observed in "low-temperature" and tempered samples, without a structural value of their own and merely marking essentially arranged and non-arranged states.

In K-Na-feldspars, optic properties have both a marking and a definite structural importance, similar to that of X-ray constants characterizing the geometry of a lattice.

The optic angle in submicroscopically twinned block crystals is reduced merely by 2 or 3° as compared with 2V in an individual. A change in the cryptoperthite composition, too, has but a slight effect on 2V: a 40% increase in the albite component content increases the optic angle by only 4 to 7°.

At the same time, in native and burnt K-feldspars, 2V changes from -84° to 0°, and on to -60° in the (010) plane; i. e., by 114°.

A comparison of 2V in samples where the distribution of Al and Si has been established by structural analysis (Spenser's "C" samples, unburnt, 2V = -44°; "sandinized", 2V = 44.8° // (010); intermediate microcline "U", 2V = -76.2°), and also with X-ray constants of samples for which the Al - Si distribution has been studied by nuclear magnetic resonance [28], shows the structural importance and the direction of change of the optic angle.

The fact that it is possible to isolate the effect of other factors on the optic angle and to determine its slight variations dependent on composition and submicroscopic twinning, makes it possible to use this angle as a criterion of the arrangement in essentially K-feldspars; this criterion requires but a slight correction for the composition and twinning.

Inasmuch as the possibility of a different distribution of Al and Si in K-feldspars is more complex than in diatomic highly symmetrical metal alloys, so that different arrangements are possible, we introduce the term, "optic arrangement", with a restricted physical meaning: i. e., the state of arrangement corresponding to the optic angle, as determined by interpolating between the extreme values of 2V known to correspond to the extreme values of the arrangement observed; the intermediate values are related merely to intermediate states of arrangement.

Optic orientation and angles of extinction in zone (010), specifically for (001) and 1 (100), are related to changes in symmetry and are therefore important in principle in the study of transformations accompanied by such changes.

A deviation of monoclinic symmetry in K-Na-feldspars, measured by angle \angle (010) \angle γ or by the extinction angle on (001), may be designated as "optic trilineity" (Δ_0), in analogy with "x-ray trilineity" (Δ_r), determined [5] from the angular distance between (131) and ($\bar{1}31$), in powdergrams.

A triclinic symmetry is the sign of a certain degree of arrangement, with the degree of trilineity dependent on the arrangement. However, arrangement is not the only factor affecting the trilineity. The latter depends also on submicroscopic twinning. A balanced submicroscopically twinned K-feldspar will have a zero total trilineity; i. e., will be pseudo-monoclinic, even when the twin individuals represent arranged maximally triclinic microclines (Δ_0 and $\Delta_r = 1.0$). Deviations from the monoclinic symmetry are associated here with unbalanced twinning.

Irrespective of the degree of arrangement in individuals, the optic trilineity may equal zero. The arrangement determines only the maximum trilineity possible in a given block crystal, which attains a limit possible for the given state of arrangement in a nontwinned homogeneous monocrystal.

Thus we have that a) the optic angle in K-feldspars is a measure of arrangement; and b) optic trilineity depends on arrangement and submicroscopic twinning.

Different behavior of optic axes A and B, in arrangement. Optic axis A in triclinic alkali feldspars (so designated in accordance with symbols adopted for plagioclase) forms an angle with the normal to (010) smaller than that made by axis B. In a "pure" arrangement, accompanied by a maximum deviation from monoclinic symmetry, optic axis B changes its position merely by an angle of up to 10°. At the same time, axis A shifts by about 38°. The position of axis B in the highest microcline is

the maximum possible for monocrystalline K-feldspars. Its shift beyond that critical point indicates the presence of a submicroscopic (or sub-X-ray) twinning. Axis A is that of the arrangement; axis B, of submicroscopic twinning.

Two types of optic and X-ray constants for K-feldspars. A redistribution of Al and Si in arrangement involves interrelated changes; viz., 1) in density and in the symmetry of texture for electron clouds; and 2) in the lattice geometry. The first is expressed in the change of γ , β , α , 2V, and optic orientation; the second, in the change of the axes and angles of the lattice. The following changes can be compared: in refractive indices α , β , γ , and the size of the lattice axes a, b, and c; optic orientation and the lattice angles α , β , and γ .

It should be noted, however, that absent among the X-ray constants is one which, like the optic angle related to γ , β , and α , could have been an intensifier of small changes in γ , b, and c.

Accordingly, two types of constants for K-feldspars can be designated by their structure: 1) triclinity, X-ray and optic; and 2) optic angle.

All constants of the first group; i.e., all X-ray and optic constants of K-feldspars, except for 2V, generally maintain the same value; they determine the triclinity as a simultaneous function of two factors, arrangement and submicroscopic (for optic constants) or sub-X-ray (for X-ray constants) twinning. The effect of this twinning overcomes the effect of arrangement. By means of the first group of constants (including the use of X-ray methods), it is impossible at the present time to isolate the effects of these two factors.

The only second group constant, 2V (without an analogue among X-ray constants), depends, in the case of K-feldspars, on the degree of arrangement, only.

The Existence of Different Structural-Optic Zones

The concept of structural-optic types. Petrographic literature carries contradictory data on the relationship of optics and the symmetry of K-feldspars. According to the "classical" view, there are only monoclinic sanidines and orthoclases and triclinic microclines. On the other hand, numerous measurements by the Fedorov method show that K-feldspars with almost any values of 2V (down to 20° and less) may have a triclinic orientation, double that of the highest microcline. Moreover, "it appears that there is no connection between 2V and the optic orientation" [4].

As pointed out before, such concepts are based on erroneous measurements. Such being the case, the question of the relationship between optics and symmetry arises again. More specifically, is "K-anorthoclase" truly triclinic? What is the limit of 2V in triclinic symmetry? Are there triclinic K-feldspars with small to intermediate 2V's, and monoclinic F-feldspars with large 2V's? How are the limits of variation for optic orientation to be determined? What is the structural significance of the combinations of optic constants observed?

On the other hand, to what extent do the X-ray data based on the triclinity measurements indicate the state of arrangement; and to what extent are they a result of sub-X-ray twinning? What, specifically, are the "intermediate K-feldspars" of J. R. Goldsmith and F. Laves [5]? Are they intermediate between sanidine and microcline or between orthoclase and microcline? Do the theoretically postulated unstable feldspars truly exist, and what are the tangible bases for their designation? Would it not be advisable perhaps to review, by applying Fourier's synthesis, the results of structural analysis for all samples used as control (T. F. Barth, F. Laves) in theoretical studies of K-feldspars?

All these, and possibly other problems, can be thoroughly systematized and considered, within the scope of material on hand, in the light of the structural-optic types.

By structural-optic types we understand those types of K-feldspars identified by arrangement (2V and X-ray and optical triclinity) related to submicroscopic and sub-X-ray twinning and to the arrangement.

Limit of variation in the properties of K-feldspar proper is established from Figures 2, 3, and 4, with points corresponding to data in Tables 1 and 2. Over a thousand determinations of optic orientation in goniometrically oriented sections, not included in Table 1, yielded values not exceeding the limits in Figure 3. Consequently, the possibility of such a deviation from the limit should at least be regarded as not demonstrated. Should such deviations have been obtained as a result of accurate measurements, they would have indicated the existence of a new variety of K-feldspar, not explainable in the light of known facts.

Figure 3 demonstrates the existence of the following structural-optic types (different combinations of 2V and triclinity):

- 1) K-feldspars with an optic angle less than about 40° in plane (010), and with the optic plane // (010), are monoclinic.
- 2) Monocrystalline K-feldspars intermediate

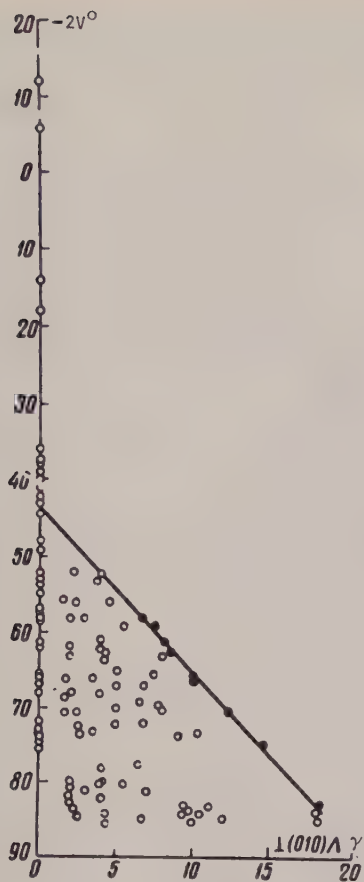


FIGURE 3. Relationship between optic orientation and the optic angle, from the measurements on essentially K-feldspars

between sanidine and microcline (their corresponding points fall on the straight line in Figure 3, connecting the points of sanidine and maximum microcline), are rare. Here belongs the unique "Taymyr II" specimen (see [17]).

3) Optically monoclinic K-feldspars form an unbroken series, up to $2V = -76^\circ$, a value which we have observed. Very small deviations (about 2°) from the monoclinic orientation have been observed as far as $2V = -84^\circ$.

4) Particular attention was paid to an accurate determination of the orientation of triclinic K-feldspars differing from intermediate microclines (of type 2).

The technique used was sufficiently sensitive for a certain identification of even small (over 1°) deviations from a monoclinic orientation. On the other hand, the accuracy of measurement ($\pm 1^\circ$) precluded mistaking the optically monoclinic crystals for triclinic, as well as an exaggeration of triclinity.

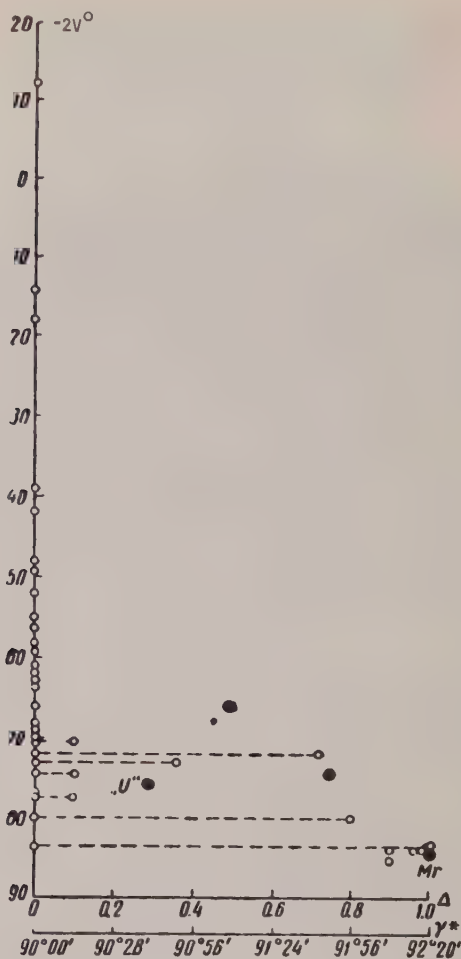


FIGURE 4. Relationship between X-ray constants: "triclinity" (Δ), angle of inverted lattice, γ^* , and the optic angles of essentially K-feldspars

Circles designate points of $2V$ and the degree of triclinity as determined from X-ray photographs by diffractometer (see Tables 1 and 2); dashes connect points for the same sample; black circles designate points for $2V$ and angles γ^* ; "U" - from data by S.W. Bailey and W.H. Taylor [27]; Mr - after F. Laves [9]; other blackened points have been obtained from Weissenberg-grams for various segments of the "Taymyr II" sample (see [17]).

Particularly convincing are observations of the deviations from monoclinic orientation, in a preparation where different positions of the indicatrix have been observed for different segments with the same crystallographic orientation.

It was taken into account, in those observations, that the block structure and bends in crystals, often present in K-feldspars, may

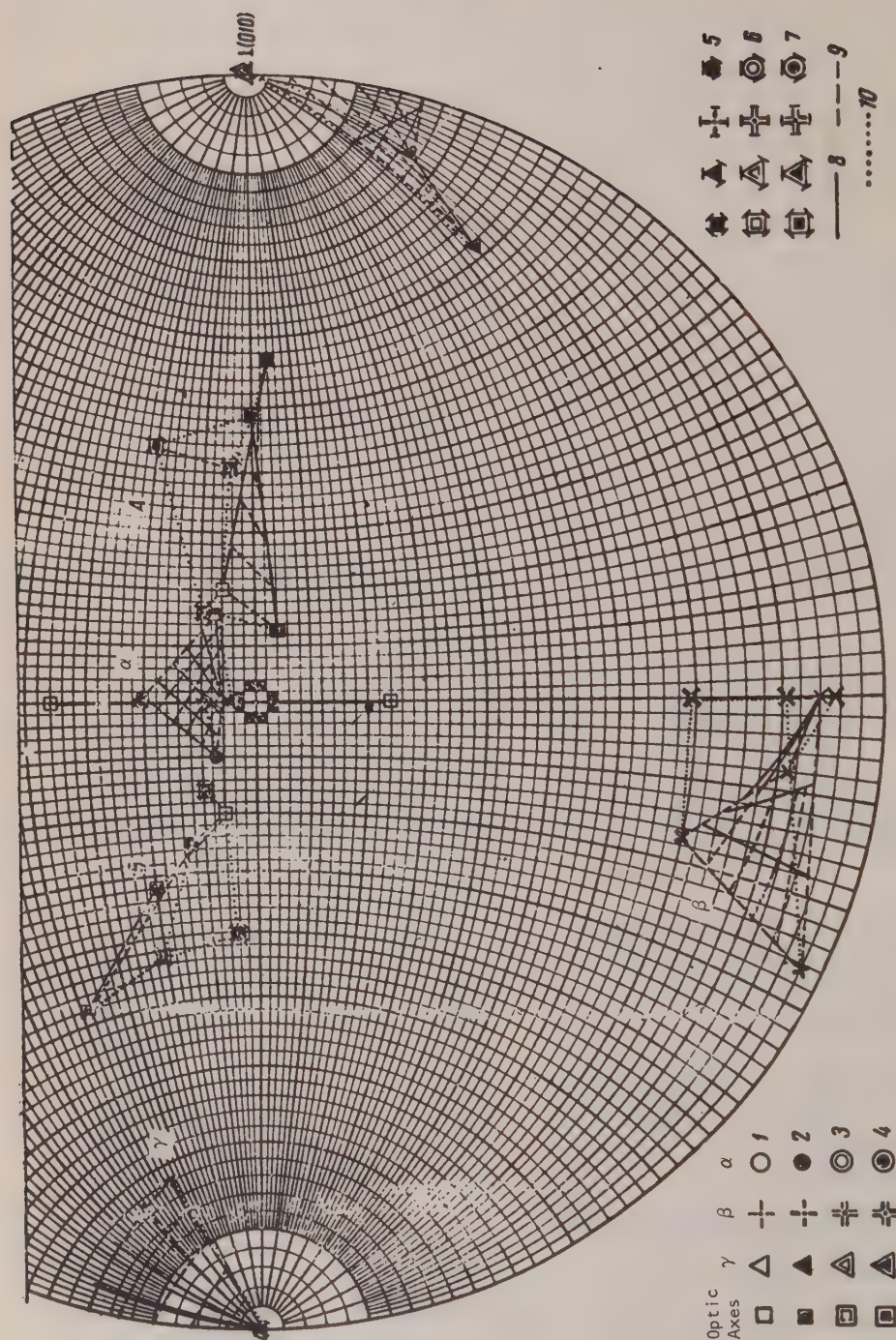


FIGURE 5. A diagram of relationship between optic orientation of K-Na-feldspars and composition, degree of arrangement, and submicroscopic twinning

1 - non-arranged K-feldspar (sanidine); 2 - arranged K-feldspar (microcline); 3 - nonarranged Na-feldspar (analcite); 4 - arranged Na-feldspar (albite). Submicroscopically twinned: 5 - microcline; 6 - analbite; 7 - albite; 8 - lines of the same composition; 9 - lines of the same degree of arrangement; 10 - lines connecting different types of submicroscopically twinned feldspars.

distort the observation results. To avoid that, the measurements were done directly at the face whose inclination was determined goniometrically; particular attention was paid to possible bends in cleavage fractures.

For higher certainty, the orientation of the optically studied sample was checked by X-ray photography. For that purpose, small fragments of 0.5 to 1.0 mm were taken from preparates (thick, uncovered slides). (This operation was comparatively simple because the sections were prepared approximately normal to the cleavage.) These fragments were oriented on the goniometric head of an X-ray camera, from the same goniometric data used in optic observations. X-ray photographs so obtained corroborated the correct setting of the sample.

The existence of monoclinic K-feldspars with the optic angles less than 40 or 50° indicates the absence of unstable sanidine, which, in turn, suggests a directed shift of Al, in the process of arrangement: first from tetrahedral positions $A = Si_2$ to $B = Si_1$, then from $B_1 = Si_1(m)$ to $B_2 = Si_1(O)$.

Figure 5 has been constructed on the basis of these measurements, calculations, and interpretations.

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BRIEF COMMUNICATIONS

THE STRUCTURE OF THE SEDIMENTARY PART OF THE MANTLE IN SOME AREAS OF THE PACIFIC, FROM REFLECTED SEISMIC WAVE DATA¹

by

S. M. Zverev

1. The Institute of the Physics of the Earth at the U. S. S. R. Academy of Sciences conducted deep seismic sounding of the earth's crust in the northwestern Pacific and adjacent basins [1, 2]. Its main purpose was to trace seismic waves from deep-seated discontinuities. As a byproduct, data on the structure of the upper level of oceanic sediments were obtained.

2. Simultaneously with the observation of deep seismic waves, special studies were made of vertical reflections from the bottom surface and sedimentary boundaries. That was done in two ways.

In exploding large charges (over 100 kg) the deep seismic sounding reflections were recorded by the channel installed in a moving ship, with a piezoelectric crystal attached to the underwater part of the ship, as a receiver. The signals were amplified and recorded in a wide frequency range; they also passed through a narrow-band filter. The filter frequency was selected to raise the effective sensitivity of the channel. Experience has shown that reflections are recorded well in frequencies of 50 to 70 cycles/sec. A seismogram so obtained is illustrated in Figure 1. Present in this and similar records, in addition to reflections from below-bottom horizons, are impulses caused by a repeated shock and possibly by a reflection from the water surface at the point of explosion.

By this method, we succeeded in obtaining

¹O stroyenii osadochnoy tolshchi nekotorykh uchastkov tikhogo okeana po dannym seysmicheskikh otrazhennykh voln.

reliable recordings of reflections from sedimentary interfaces, at oceanic depths of 5 to 6 km. In deep sinks, the reflections weakened to where the effective sensitivity of this method was inadequate for their reliable registration.

The explosion and reception conditions were sufficiently uniform for a correlation of the behavior of reflecting surfaces over large areas.

3. During the brief halts, reflections were registered with a multichannel floating installation, similar to those used in marine exploration for oil. An installation with 6 hydrophones at a 50 m interval was used in the 1958 work. Their spread was achieved by the wind drift of the ship. Seismic observations were carried on simultaneously with oceanographic soundings. The charges were 1 to 5 kg, attached to rubber floats. Amplifiers similar to the seismic exploration type were used. The frequency of registered reflections was 30 to 35 cycles. A sample seismogram is shown in Figure 2. Automatic control of amplification and mixing was not used, so that it was possible to utilize the dynamic characteristics of seismograms.

4. The multichannel installation had a higher effective sensitivity than the single-channel one, on the moving ship. Its recordings were free of interference by repeated shocks. The application of correlation criteria increased the reliability of the reflection identification; accordingly, the multichannel seismograms were taken as basic for tying up the reflections obtained from a moving ship.

Differences in the technique of explosion and recording have led to appreciable differences in the absolute timing of reflections from the same horizons but registered by different devices. After a correction for the ship movement and for the explosion depth, this difference was considerably smaller; still, the reflection time from the bottom, as registered by the multichannel installation, systematically exceeded by 0.01 to 0.7 sec the corrected time recorded from the moving ship. This is

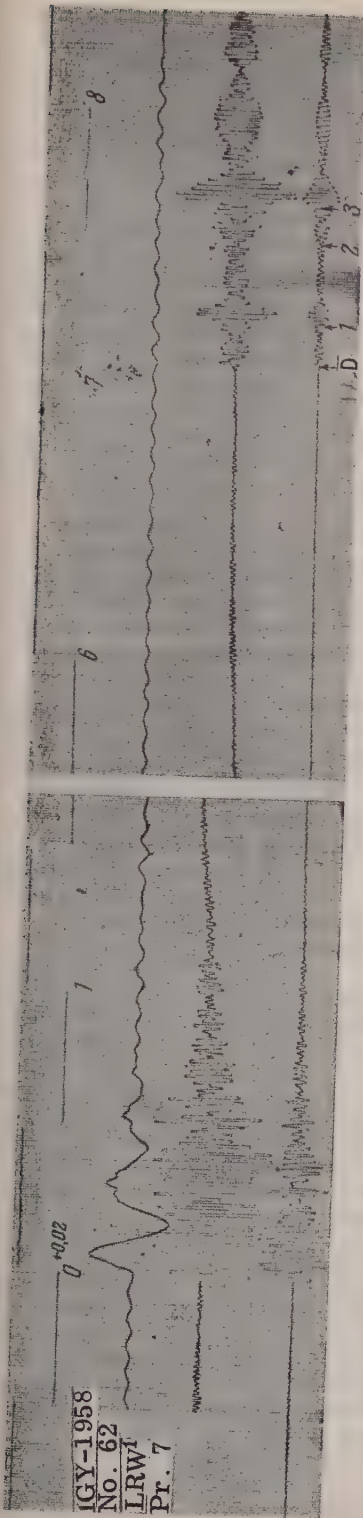


FIGURE 1. Seismogram of reflections from the bottom surface (D) and sedimentary boundaries [1, 2, 3], obtained on a moving ship; tracks 3 and 4 from the top - records with different amplifications.

¹Longitudinal Reflected Wave.

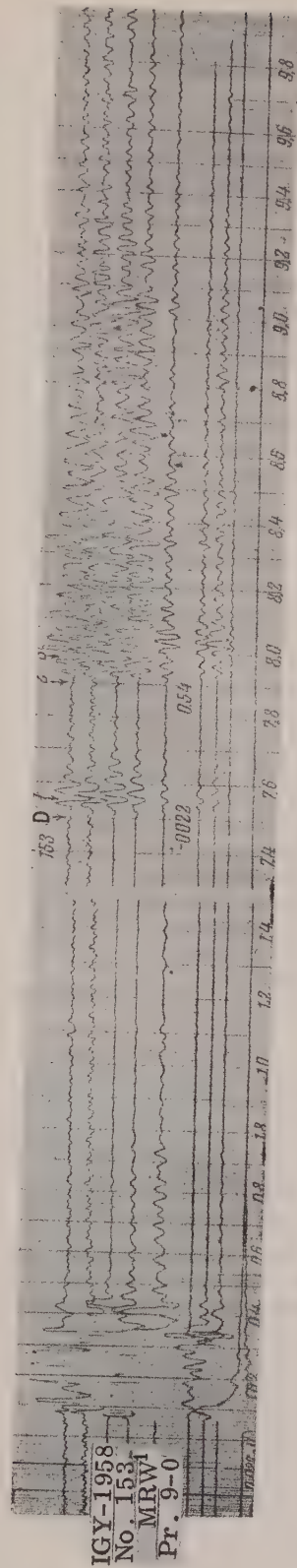


FIGURE 2. Seismogram of reflections from the bottom surface (D) and sedimentary boundaries [1, 2, 3] obtained with a multichannel installation.

Tracks 1-5 from the top - recordings by the several hydrophones in the multiple installation. Tracks 6 through 8 - recordings by hydrophones 1, 3, and 5, of a lower sensitivity.

¹Multiple Reflected Wave.

possibly due to an uneven penetration of sediments by waves of different frequency (low frequency waves penetrate deepest).

By means of the multichannel installation, it was possible to register reflections from sedimentary boundaries, in the oceanic bed as well as in a deep trough, and in some places on the shelf.

5. It is to be noted that cross-sections obtained from vertical reflections are subject to distortion by the phenomena of multiple reflection of waves in layers; also by interference of waves with a length commensurable with the thickness of layers; and by other complicating phenomena which take place in seismic exploration with reflected waves. In our study of relatively thin sediments under a great depth of water, and in the absence of sufficiently long time-distance curves, an identification of all these interfering phenomena was even more difficult than in standard seismic exploration. Because of that, the profiles so obtained should be regarded merely as the simplest interpretative variant only partly corresponding to the true picture.

6. Observations of vertical reflections were indicated in all deep seismic sounding profiles run in 1958. We shall consider data from the Pacific profiles, near Kamchatka and the Komandorskiy Islands. An index map showing this area is given by Ye. I. Halperin, A. V. Goryachev, and S. M. Zverev [2]. Profiles 7 and 9 cross the continental shelf at Kamchatka, proceed across the deep Kurile-Kamchatka trough, and on to the deep Pacific basin, as much as 500 km offshore. Profile 8 crosses the shelf at the Komandorskiy Islands, the Aleutian deep trough, and ends at the Pacific marginal swell, about 150 km before the ocean-bottom plain.

Cross-sections along these profiles are presented in Figures 3, 4, and 5. The time of vertical reflection is indicated by dashes whose length, proportional to the amplitude of reflection on seismograms, characterizes the relative reflecting capacity of the corresponding boundary. Data obtained with multichannel installations are presented separately.

7. This article discusses the cross-sections of sedimentary rocks, expressed not in the thickness of their layers but in the arrival time of their vertical reflections. This is because the velocity determination in sediments, with the described technique for observing the reflected waves and vertical reflections, is quite complicated, so that despite the few data on hand, it cannot be regarded as final. However, the data cited are quite adequate for a comparative geologic interpretation. In addition, it is possible to indicate approximately the possible limits of velocities in sediments and so to obtain the approximate thicknesses.

8. The records of multiple reflections from sediments, obtained on seismograms, as well as the values of the reflection factors in a vertical path, as computed from the records of vertical reflections of various multiplicities, were used in estimating the velocity in sediments. An analysis of amplitude curves for multiple reflection water waves, assuming the segments of the amplitude growth to be related to critical angles [3], has yielded in one of the ocean-bottom profiles presented below, a value of 1.62 to 1.67 km/sec² for velocities in the near-bottom layer; and of 1.89 to 0.91 km/sec for a deeper layer. The fair coincidence of the experimental and the corresponding theoretical data (Figure 6) permits the assumption that the second velocity represented transverse waves. Accordingly, we obtain a value for the velocity for longitudinal waves in a deeper sedimentary layer, on the order of 3.2 to 3.5 km/sec.

From the ratio of amplitudes for vertical reflections of single and double reflections, the following values of the reflection factor were obtained in the profile: for the bottom surface, 0.17; for the sharpest boundary, 0.58; and for the deepest reflecting boundary, 0.34.

A computation with Rayleigh formulas for the vertical reflection factor has shown that the following velocities V and densities ρ in sedimentary layers are in accord with these data:

a) ocean-bottom sediments (unconsolidated):	$V = 1.65 \text{ km/sec}$ $\rho = 1.35 \text{ gm/cm}^3$
b) deeper sediments (consolidated):	$V = 3.5 \text{ km/sec}$ $\rho = 2.5 \text{ gm/cm}^3$
c) base of sediments (crystalline rocks):	$V = 6.4 \text{ km/sec}$ $\rho = 2.8 \text{ gm/cm}^3$

As we shall see below, the 1.65 km/sec layer on the ocean bottom is usually 300 to 600 m thick; the 3.5 km/sec layer is about 130 to 300 m thick. This gives an approximate average velocity of about 2 km/sec for the entire sedimentary section.

All these velocity estimates are quite approximate, with not all of the data extant having been used. More specifically, the presence of a thin compact sedimentary layer with a velocity of about 3.5 km/sec cannot be regarded as definitely established. A more accurate determination of velocities in sediments requires additional study.

²Here and elsewhere in this article, we do not consider the thin layer (probably about 5 to 15 m thick) at the very surface of the bottom. A weak reflection from such a layer can be seen in the seismogram of Figure 2.

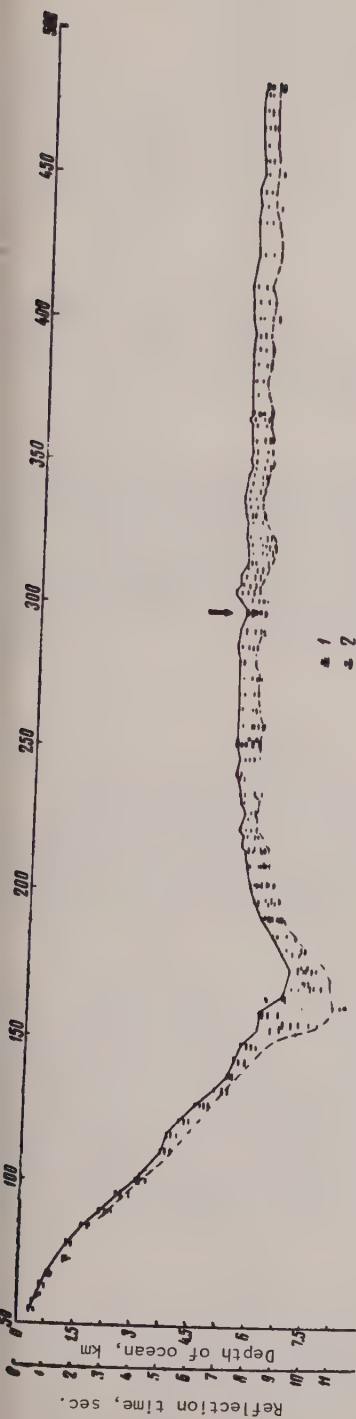


FIGURE 3. Arrival time for vertical reflection in profile 9 crossing the Kurile-Kamchatka deep trough, from the southern tip of Kamchatka

1 - time obtained from observations on a moving ship; 2 - same, for the multichannel installation; 3 - points on profile with an abrupt change in the arrival time for reflections from the bottom and from sedimentary boundaries.

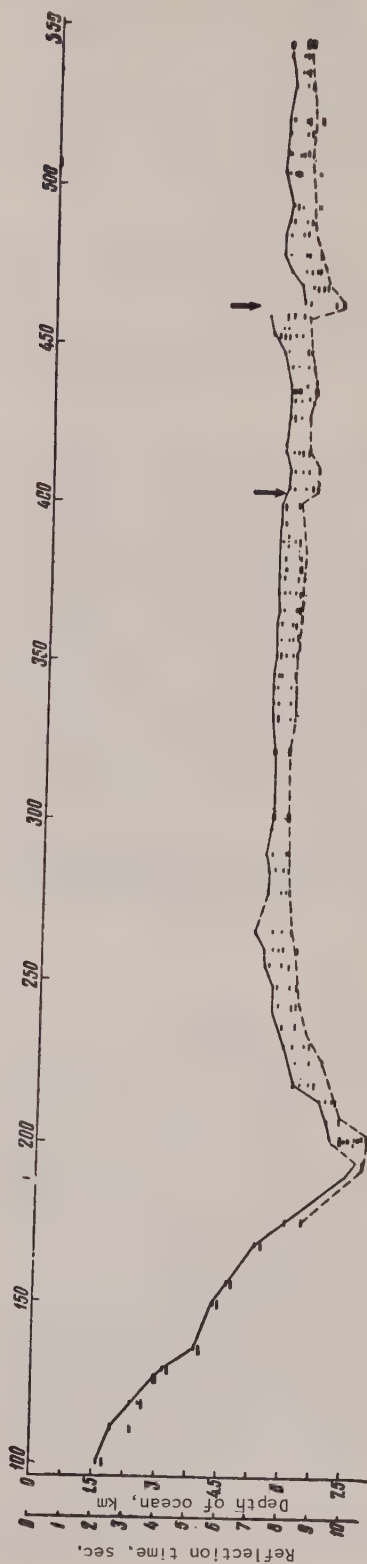


FIGURE 4. Arrival time for vertical reflections in profile 7 crossing the Kurile-Kamchatka deep trough, from the area of Petropavlovsk-on-Kamchatka, Symbols the same as in Figure 3.

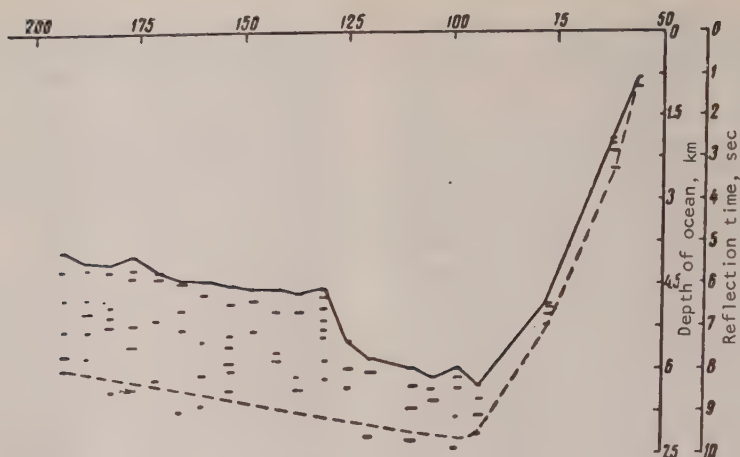


FIGURE 5. Arrival time for vertical reflections in profile 8, crossing the Aleutian trough from the Komandorskiy Islands.

9. In all three profiles, the nature of sediments changes abruptly, in crossing the deep trough. On the ocean side, they are definitely and consistently stratified. The amplitude of bottom reflections is usually 2 to 4 times smaller than that for later reflections, which suggests that the bottom here does not present as sharp a boundary as deeper sedimentary interfaces. Between the trough and the land, the stratification is less conspicuous; the reflecting properties of the bottom are quite variable, being on the whole 2 or 3 times higher than in the ocean-bottom sediments. This suggests a higher density of propagation velocity in the ocean bottom rocks and a rapid change in their composition.

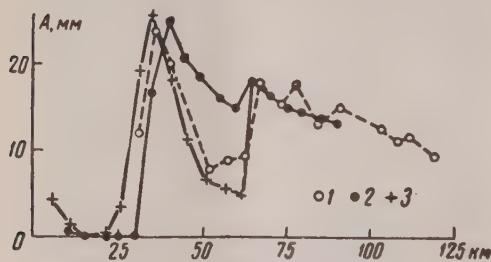


FIGURE 6. Correlation of experimental and calculated amplitude graphs for triple reflection waves in ocean water, for determining the propagation velocity in sediments

1 - experimental data; 2 - calculated data for the first critical angle (second maximum in graph corresponds to the near-bottom layer velocity of about 1.65 km/sec; first maximum corresponds to the velocity of longitudinal waves in a deeper sedimentary layer; i.e., about 1.90 km/sec); 3 - calculated data for the second critical angle (first maximum corresponds to the velocity of transverse waves in a deeper sedimentary layer; i.e., about 1.90 km/sec).

10. Turning to profile 9, passing southeast of the south tip of Kamchatka, the western slope of the deep trough carries, generally speaking, three reflection horizons with time intervals Δt between the first and the last reflections, of 0.2 to 0.4 sec (Figure 3). Identifiable in the deep trough are 5 to 6 reflection horizons with Δt up to 1.5 or 1.8 sec.

From 2 to 5 horizons have been observed in the marginal oceanic swell touching the southeast slope of the trough. Near the trough, Δt changes within a fairly wide range of 0.2 to 0.9 sec, which suggests complex relief underneath the sedimentary mantle. About 130 km away from the trough axis, there is an abrupt change in the bottom relief and a displacement of the reflecting areas — evidence of faulting in the bedrock. Farther away from the trough axis, 3 to 4 reflections have been observed; the reflecting relief becomes more subdued; and Δt between the reflections gradually drops to 0.4 to 0.7 sec. Assuming an average velocity in sediments, of 2 km/sec, we obtain a thickness of 1.5 to 1.8 km for stratified sediments in the trough; 0.2 to 0.9 km/sec in the ocean-bottom beds near the trough; and 0.4 to 0.7 km in the southeastern segment of the profile. Within the trough, the figure of about 2 km possibly is not representative of the entire stratified sedimentary section because it appears that no reflections have been obtained from its base.

Within the northwestern slope, the total thickness of rocks down to the "granite" cannot be inferred from material on hand. It can only be stated that the thickness of approximately horizontally stratified sediments, yielding appreciable seismic reflections, is not over 300 m. According to deep seismic sounding data, they too are underlain by low-velocity rocks. The absence of reflections

suggests a complex structure of the sedimentary mantle.

11. The cross-section along profile 7 (Figure 4) is similar on the whole to that along profile 9, with sediments near the southeastern slope of the trough having a more subdued relief. Here, Δt ranges from 0.5 to 0.9 sec (thickness, 0.5 to 0.9 km). A normal fault cutting all sediments and expressed in a scarp of about 800 m in bottom relief is present at about 250 km from the trough axis; a smaller fault passes 50 km closer to the axis. The thickness of sediments at the southeastern end of the profile is 0.4 to 0.7 km.

12. We turn now to the profile running southeast of the Komandorskiy Islands, across the Aleutian trough (Figure 5). These data are less reliable because bad weather made it impossible to use the multichannel installation.

On the shelf side, there are 2 to 4 reflection horizons with Δt of no more than 0.5 sec. The number of these horizons increases abruptly to eight and more, in the trough area and south of there, with Δt 2.0 to 2.5 sec; i. e., 2 to 3 times greater than in the ocean-bottom beds, in profiles 7 and 9. This indicates that the sedimentary section in the Pacific marginal swell area near the Aleutian trough is at least 2.0 to 2.5 km thick.

SUMMARY

1. A deep trough separates a horizontally stratified, comparatively unaltered sedimentary section at the ocean bottom, from the complex shelf section. This bottom section thickens gradually away from the trough.

2. The surface of rocks underlying the sedimentary bottom section, within the marginal swell of the Kurile-Kamchatka trough, as a complex relief smoothed down on the whole by sedimentation. Away from the trough, relief at the sedimentary base becomes more subdued. Normal faults with large displacements, and cutting all sediments, form scarps in the bottom relief, 150 to 250 km away from the trough axis. These faults probably originated in comparatively recent tectonic movements which affected the ocean bottom at the Kurile-Kamchatka trough.

3. Sediments within the marginal Pacific swell at the Aleutian trough are 2 or 3 times thicker than in the ocean-bottom beds near the Kurile-Kamchatka trough. This suggests essentially different sedimentary conditions in the outer parts of the Kurile-Kamchatka and Aleutian troughs.

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CALCIC AEGIRINE — A MEMBER OF THE ISOMORPHIC MONOCLINIC PYROXENE SERIES³

by

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Studies of crystalline rocks of the Iyengrian Archean series from the productive Fedorov horizon of south Yakutia uncovered dark-green, almost black monoclinic pyroxene, as a rock-forming mineral in various plagioclase schists.

These schists, changing gradually to gneisses of a similar composition, form thin to thick-bedded members consisting of rapidly alternating beds and lentils of pyroxene-plagioclase, amphibole-pyroxene-plagioclase, and amphibole-plagioclase schists. The areas of migmatization and of granite and granite-pegmatite veins (coming out of locally developed alaskite) in these crystalline schists show green-yellow epidote and dark-brown andradite, in places in considerable amount. Externally, and

³Kal'tsioegirin — chlen izomorfnoy ryady monoklinnykh piroksenov.

depending on their mineral composition, these plagioclase schists form a light- to dark-gray fine-grained, thin-banded (or dense, horn-felsic) rock with vein-like spots of pyrite and pyrrhotite, occasional quartz grains, common accessory apatite and sphene, and rare fluorite. Pyroxenes are replaced locally and partially (on the periphery and in fractures) by blue-green hornblende, with phlogopite and chlorite appearing near the granite injections (on hornblende and diopside); isolated grains and mosaic pockets of orthite occur in partially skarned (hornfelsic) segments in epidote.

Relict grains of calcite indicating the calcareous nature of the original sedimentary rocks are fairly common in these rocks.

Plagioclase is usually andesine (No. 32-37), becoming more acid in migmatization zones (Pl No. 10-16); as an effect of granite injections, it is partially replaced by scapolite.

Monoclinic pyroxene occurs in fine to relatively coarse (up to 4 mm long) short prismatic crystals, octagonal in cross-section and with a typical prismatic cleavage almost at a right angle; also in irregular, slightly elongated grains, most commonly as light-green and non-pleochroic diopside ($C\gamma = 38$ to 40° ; $+2V = 52$ to 56°); in a number of layers and lenticular segments, however (especially those with magnetite inclusions), it is a dark-colored pyroxene, strongly resembling aegirine in the intensity and nature of its coloring.

Microscopic study of numerous hand specimens from different areas has shown that γ of the dark-green pyroxene is yellow green to green-yellow; β is emerald green to dark blue-green; and α is deep apple-green. The absorption, $\beta \geq \alpha > \gamma$; $C\gamma$ varies from 56 to 64° ; and $2V$ is positive and varies from 60 to 74° .

A perfectly pure dark-green pyroxene was carefully isolated by K. L. Grachev from a partially skarned (garnetized) schist and analyzed by O. V. Uranova, in 1957.

A conversion of the chemical analysis of the Aldan mineral for its component "molecules" yields the only possible result, which is given in Table 2, Column 1.

Definitely present along with diopside and hedenbergite, Cermak's and sodic aegirine "molecules" is a considerable amount of the calcic aegirine molecule, $\text{Ca}_{0.5}\text{Fe}^{+++}\text{Si}_2\text{O}_6$, where calcium replaces the sodium of the aegirine ($\text{Ca} \rightarrow 2\text{Na}$).

A similar pyroxene was encountered by S. S. Smirnov [2] and D. S. Korzhinskiy [1], in the Baykal region, in alkali granites and pegmatites; also in metasomatic rocks saturated with calcium.

Table 1

Chemical Composition of Calcic Aegirine - Diopside - Augite from Archean Rocks of Aldan and the Baykal Region

Components	1		2	
	weight %	molec. number	weight %	molec. number
SiO_2	45.29	755	51.07	851
TiO_2	None	—	0.32	4
Al_2O_3	14.25	140	2.03	20
Fe_2O_3	10.26	64	8.63	54
FeO	5.27	73	7.30	101
MnO	0.33	5	0.71	10
MgO	3.40	84	8.06	200
CaO	20.46	366	21.63	401
Na_2O	0.42	7	0.32	5
K_2O	None	—	0.12	1
P_2O_5	"	—	—	—
F	Trace	—	None	—
Cl	"	—	"	—
CO_2	None	—	"	—
S	"	—	"	—
H_2O^+	0.05	—	0.19	—
H_2O^-	0.29	—	—	—
Total	100.02	—	100.38	—

¹ Dark-green pyroxene from skarn crystalline schists in migmatite zone. Aldan, specimen 149/55. Tayezhoye: $2V = +72^\circ$; $C\gamma = 60^\circ$; $\gamma = 1.737 \pm 0.002$; $\alpha = 1.712 \pm 0.002$; $\beta = 1.721$ (as calculated); $\gamma - \alpha = 0.025$.

² Pyroxene from a pegmatite vein in Zilberman mine, Transbaykal region, after S. S. Smirnov [2]. Ye. Ya. Rode, Analyst (Labor. of Mining Institute, 1925).

Table 2

Molecular Composition of Calcic Aegirine - Diopside - Augite

Chemical formula	1	2
$\text{CaMgSi}_2\text{O}_6$	23.0	45.7
$\text{Ca}(\text{Fe}, \text{Mn})\text{Si}_2\text{O}_6$	21.9	5.2
$\text{Ca}_{0.5}\text{Fe}^{+++}\text{Si}_2\text{O}_6$	23.4	20.4
$\text{NaFe}^{+++}\text{Si}_2\text{O}_6$	2.9	2.7
$\text{CaFe}_3^{+++}\text{SiO}_6$	—	0.7
$\text{CaAl}_2\text{SiO}_6$	28.8	4.6
$\text{Ca}_2\text{Si}_2\text{O}_6$	—	0.7
Total	100.0	100.0

Black-green pyroxene in pyroxene-scapolite rocks of phlogopite deposits display the following pleochroism: β deep blue-green $\geq \alpha$ light yellow-green. In the same rocks, the following constants were measured by N. M. Keppen: $2V = +70^\circ$; $\gamma = 50^\circ$; and $\gamma - \alpha = 0.026$.

The chemical composition of deeply colored pyroxene of this type, from a pegmatite vein, is given in Table 1, analysis 2, after S. S. Smirnov [2]. A conversion to component pyroxene "molecules" was done by D. S. Korzhinskiy [1] and is presented in Table 2, Column 2.

A microscopic study has revealed, as suggested by S. S. Smirnov, the clean-cut zonation in pyroxene grains. Their marginal parts, believed by him to be diopside-aegirine-augite, are vivid grass-green with strong pleochroism and an aegirine absorption; toward the central part of the grain, the color becomes much less intense and finally fades away, as in diopside. The $2V$ in pyroxenes varies from $+64$ to $+82^\circ$; γ from 45 to 67° ; and $\gamma - \alpha$ $0.025 - 0.027$.

The maximum values of $2V$ and the extinction angle γ correspond, as in Aldan, to the deepest colored grains and parts of grains with distinct pleochroism: γ , bottle-brown; β , grass-green with a slight brownish tint; and α , pure grass-green. The absorption formula is $\gamma < \beta \geq \alpha$.

Both dark-green pyroxenites, from Aldan and the Transbaykal region, contain small amounts of Na_2O but much Fe_2O_3 ; their CaO content is about equally high in both; a relative excess of

MgO and SiO_2 in Transbaykalian pyroxene is compensated for by an excess of Al_2O_3 in the Aldan pyroxene ($\text{MgSi} - \text{Al}_2$ isomorphism).

The aegirine-like behavior of the two pyroxenes appears to be due (as previously noted by D. S. Korzhinskiy) to the considerable presence in them of the $\text{Ca}_{0.5}\text{Fe}^{+++}\text{Si}_2\text{O}_6$ molecule, which we call here calcic aegirine.

The Aldan pyroxene is high in Cermak's molecule (about 30%) but its diopside-hedenbergite molecular content (45%) is much lower than in Transbaykalian pyroxene (71%). Thus, the first is a calcic aegirine - augite, while the second is calcic aegirine - diopside.

That our south Yakutian pyroxene (specimen 149/55) belongs to the diopside-augite-aegirine series is confirmed by a Debye powder pattern (Table 3) obtained by G. V. Golubev, in 1959 (iron radiation, $2R = 57.3$ mm; exposure, 1 hr 10 min; in brackets, β radiation).

The preferential development of dark-colored pleochroic pyroxenes with calcic aegirine in Aldanian rocks containing magnetite and calcite indicates a saturation in calcium and the participation of magnetite iron in the recrystallization of pyroxenites (under conditions of recurrent metamorphism and accompanying metasomatic phenomena). However pyroxenes of that type could have been formed as primary minerals, as well, at the expense of iron and calcium-rich sedimentary rocks, in a way similar to the formation of diopside out of rocks with little or no iron.

Table 3

Interplanar Distances for the Aldan Calcic Aegirine - Augite

<i>N</i>	<i>l</i>	<i>d_{hkl}</i>	<i>N</i>	<i>l</i>	<i>d_{hkl}</i>	<i>N</i>	<i>l</i>	<i>d_{hkl}</i>
1	4 b.&d.*	6.53	16	1	(1.792)	32	3	1.257
2	4	4.48	17	5	1.761	33	1	1.242
3	1	4.05	18	1	1.688	34	3-4	1.221
4	1	3.59	19	9	1.629	35	2 b.&d.*	(1.183)
5	5	(3.24)	20	2	(1.571)	36	1/2	1.152
6	10	2.97	21	2	1.560	37	1/2	(1.119)
7	3-2	2.88	22	2	1.530	38	1/2	(1.103)
8	1	2.81	23	3	1.515	39	10	1.078
9	10	2.53	24	1	1.497	40	7	1.070
10	2	(2.32)	25	1	very weakly lined	41	6 b.&d.*	1.051
11	2	(2.22)	26	7	1.437	42	4	1.034
12	6	2.12	27	4	(1.414)	43	6	1.020
13	6-4	2.01	28	6	1.339	44	4	1.013
14	1	(1.937)	29	1	1.325	45	4	1.007
15	1	1.840	30	6	1.289	46	5	0.998
			31	3	1.276			

*Broad and diffuse.

It is of interest that the garnetization of these Aldanian calcic aegirine - augite pyroxenes was accompanied by the development of hystero-genetic fibers and grains of magnetite (and quartz) as well as of fine magnetite dust, in the migmatization zone.

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METHODS

APPLICATION OF A STAINING METHOD IN THE MINERAL ANALYSIS OF CLAYS AND THE PRACTICAL UTILIZATION OF COLOR REACTION OF CLAYS WITH ORGANIC PIGMENTS^{1,2}

by

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In recent years, the Laboratory of Physical Study Methods at the All-Union Institute of Hydrogeology and Engineering Geology (VSEINGEO) has carried on comprehensive mineral studies of clays, including the staining method, worked out theoretically by N. Ye. Vedeneyeva and now practiced by M. F. Vikulova and M. A. Ratayev.

This method is widely used in the Soviet Union for mineral analysis of clay rocks, in industrial and research organizations. Its use is stimulated by the necessity for a simple and rapid means of determining the mineral composition of clay, often under field conditions. Its further development, in both the theoretical substantiation of the adsorption process by clay minerals and in subdividing barren argillaceous sequences, has often resulted in "anomalous" coloration of clays in visual and spectrophotometric color registration. For this reason, its authors deemed necessary further and thorough study of pigment adsorption by various minerals and clays and of conditions modifying such adsorption. The specific and immediate tasks of this study are set forth in a joint work by N. Ye. Vedeneyeva and M. F. Vikulova [4].

Guided by these tasks, the VSEINGEO

Laboratory of Physical Study Methods, has carried on experimental research in the development of the staining method through the use of modern spectrophotometric registering devices with a monochromator. It has been determined that the recording spectrophotometer SF-2M is suitable in the work with colored clay suspensions in water. At the same time, it turned out that the adsorption process of the blue methylene dye (MB) was more complex than hitherto believed; that was expressed practically in the presence of many more adsorption maxima, as compared with N. Ye. Vedeneyeva's data obtained mostly with the FM photometer. Our attempt to interpret this new experimental material in the identification of clay minerals forced us to review the familiar premises of N. Ye. Vedeneyeva's experimental methods and to introduce a quantitative factor, the amount of pigment to a unit of active surface, dependent on the sorptive capacity of the material studied. In that way, we obtained correlative results for various mineral adsorbents; it also became clear that the new data could not be interpreted by the N. Ye. Vedeneyeva hypothesis of a "double adsorption" of MB dye by clay minerals; there remained many unexplained facts.

According to A. V. Kiselev [11], "many facts have been accumulated in recent time, showing that, even at common temperatures, the reactions are by no means merely physical. Under the same conditions and for the same adsorbent, there often prevails a complex association of mutually complementary phenomena which may lead not only to physical electrostatic and dispersion reactions but to the formation of surface complexes and even of new surface chemical compounds, as a result of chemical reactions with the adsorbent. Consequently, A. V. Kiselev believes that "an adsorbent cannot be regarded as a "dead" substance not undergoing any further changes in the adsorption process itself. That substance is chemically "alive" and always subject to change".

The principal present trend in the field of adsorption phenomena is the study of surface chemical reactions and chemical compounds. It was in that light that we considered the

¹ O primeneni "metoda krasiteley" v mineralogicheskoy analize glin i puti prakticheskogo ispol'zovaniya tsvetnykh reaktsiy glin s organicheskimi krasitel'yami.

² Paper read at the expanded Bureau of the S.S.R., Academy of Sciences Commission on the study of clays, May 7, 1960.

experimental data of spectrophotometric analyses for MB water suspensions of aluminosilicates and their dehydration products. It turned out that certain facts, unexplainable by the "double adsorption" theory of N. Ye. Vedeneyeva, can be understood in the light of our hypothesis of surface reactions between aluminosilicates and the basic MB dye. This hypothesis of ours calls for a further and comprehensive study.

Theoretical premises for the staining method were developed by N. Ye. Vedeneyeva, through experimentation with the adsorption of MB dye by clay minerals in a water medium. Inasmuch as the authors of the method recommend this dye as basic, we directed our main effort to the study of the MB adsorption, without considering benzidine and chrysoidine (in a salt of hydrochloric acid), although they were also recommended as auxiliaries. Of interest in this connection are works of J. Bloch, J. Charbonnelle, and F. Kayser [20] and V. I. Yezdakov [8] who have demonstrated the dependence of benzidine staining of clay minerals on the presence of iron oxide ions in the mineral lattice and in the dispersion medium.

The Experimental Setup

A registering spectrophotometer SF-2m was used in obtaining continuous spectrum absorption curves for light reflected from stained surfaces of highly dispersed minerals in 1% water suspension or in stained powder. The work was done with a 2 mμ resolution, at a maximum recording speed. The quantity measured was the light reflection in percent, as compared with the unstained sample; i. e., the quantity $(I/I^0) \cdot 100$. In this particular study, the use of a logarithmic scale gives inferior results. For a better correlation of spectrophotometric results with the work of N. Ye. Vedeneyeva, i. e., in order to obtain a maximum light absorption, quantities $[(J_0 - J)/J_0] \cdot 100$ were used in constructing the spectrum curve.

The MB adsorbing capacity was first determined for all adsorbents under study, by the R. Robertson and R. Ward method [23], with minor simplifications of our own [13]. The staining of adsorbents was done with increasing doses of the MB dye: 0.1, 0.5, 1.0, 5, 10, 30, 50, 70 and 100% of E (total adsorption capacity). The quantitative evaluation from spectrum curves was done in order to determine regularities in the adsorption of dye for the several crystallographic elements of clay minerals (basal active faces and in places of broken bond). Selected for that purpose was quantity J_{570}/J_{680} ; i. e., the ratio of the light absorption intensities by MB in clay minerals with absorption maxima of 570 and 680 mμ.

The accumulated spectrophotometric data

have demonstrated a definite mutual relationship between the 570 mμ maximum, on the one hand, and the 620, 640, 660, and 680 mμ, on the other. The 680 mμ maximum is the least affected by overlapping effects of the adjacent maxima because it is the extreme long-wave maximum (not counting the 730 mμ maximum which appears later on and whose intensity grows with time). The color effect of KCl was evaluated from spectrophotometric data with the relative quantity J_{680}/J_{680}^{KCl} which is the intensity ratio of the 680 mμ light absorption maximum on the two spectrum curves of adsorbed dye: MB and MB + KCl. In the absence of the color effect, this quantity is equal or nearly equal to one. When the effect is present, this quantity is greater than 1.

Various clay and other silicate minerals were used for adsorbents, in fractions $< 1 \mu$: 1) "monomineral" clays from known deposits; 2) natural polymineral clays; and 3) synthetic mixtures of clay and non-clay minerals. All samples were studied by a combination of methods: X-ray (under the direction of Ts. M. Raytburd), thermal, electron-microscopic, and chemical analyses. The following were used as standard minerals:

1. Kaolinite; fraction $< 1 \mu$ of the Prosyanyov, Glukhovets, and Kyshtym kaolin and of the Latninsk clay.
2. Montmorillonite; fraction < 1 of the Oglanlin bentonite; gumbo clay (Georgian S. S. R.); fuller's earth (Crimea); askangel (Georgian S. S. R.); and pink montmorillonite (Transbaykal region, Zavityaya).
3. Hydromica; fraction $< 1 \mu$ of top loam from the spotty tundra, the Vorkutskaya oblast (sample 32); South Caspian oozes (obtained from the Hydrogeology Laboratory, the U. S. S. R. Academy of Sciences).
4. Monothermite; fraction $< 1 m$ of the Chasov'yarsk monothermite clay.
5. Chlorite; Kompirsay village, the Kazakh S. S. R.
6. Muscovite; Mamsk deposit, Irkutskaya Oblast.
7. Sepiolite; Novaya Akkermanovka, the Urals.
8. Talc; Onot River, East Siberia.
9. Pyrophyllite; Zaved Bakhgovsk, the Urals.

A total of 30 samples of minerals and natural mixtures was used. Only part of the experimental data are cited in this work.

The Main Premises of the N. Ye. Vedeneyeva Staining Method". Facts Unexplainable in the Light of the Double Adsorption Hypothesis

The basis of N. Ye. Vedeneyeva's "staining method" is the hypothesis of double adsorption of certain basic dyes by clay minerals. A prerequisite of such adsorption is a fixed distance between active atoms of the adsorbent and the pigment molecule, possible, according to N. Ye. Vedeneyeva, on basal faces of clay minerals. The principal diagnostic criteria of this method are the "stability" of the double adsorption of the pigment and a "fast bond" of its cations and the adsorbent. According to N. Ye. Vedeneyeva, the "stability" of double adsorption is the capacity of a doubly-adsorbed pigment cation to withstand the withdrawal of the second active cation atom, while maintaining the ion bond of the pigment cation with the surface, with the withdrawal releasing the cation from that double bond and to the state of a single ion bond. A diagnostic criterion in N. Ye. Vedeneyeva's spectrophotometric analysis is the position of maxima of light absorption, in the 560 to 570 m μ range. It was assumed conditionally that the maximum light absorption for 560 to 580 m μ characterizes a monomeric pigment cation, in double bond at the surface of clay particles, while the 680 m μ maximum characterizes pigment cation in a single-point bond with that surface.

On that basis, N. Ye. Vedeneyeva [3] divided the clay minerals into three principal and two transitional types, as follows:

1. Type K. A stable double adsorption; weak ion bond (kaolinite and hydromica).
2. Type M. Unstable double adsorption; strong ion bond (montmorillonite).
3. Type B. Double adsorption missing; strong ion bond (typical beidellite).
4. Type KM (hydromicas, monothermite).
5. Type MB (non-typical montmorillonite and beidellite).

The main techniques of N. Ye. Vedeneyeva's "staining method" are the introduction of K-ions into a colored water suspension; and its heating and vaporization. According to the Vedeneyeva hypothesis, the introduction of K-ions brings about a "stability" and the entire phenomenon of double adsorption. The color effect of heating the MB dyed clay suspension in water characterizes both the "stability" of double adsorption and the "bond strength" of cations and the adsorbent. The effect of drying up the colored clay mineral suspension in water is a characteristic of the "strength" of the cation bond of the MB dye with the adsorbent.

Some of our own experimental results cannot be interpreted in the light of that theory. Among these are the following:

1. Spectrum curves for the MB dye vary in the same way, qualitatively, in various layer-structure aluminosilicates with active basal faces (montmorillonite, kaolinite, hydromica, mica, chlorite, monothermite, etc.), in a water medium, at the introduction of K-ions in the colored suspension (see Figure 1).

With comparatively small doses of the dye (compared with the adsorption capacity E), color effects have been observed for all those minerals. According to the "double adsorption" theory, however, kaolinite and hydromica have a stable double adsorption, and there should be no color effect of K-ions.

2. Color effects are qualitatively similar at the introduction of K-ions in colored water suspensions of the minerals named above, in comparing the suspensions colored by relatively similar doses of MB.

Table 1 presents the results of corresponding conversion of photometric data. It shows that the color effect of K-ions decreases as the relative dose of the MB dye increases. For the same relative values of the MB dye, the quantity J₆₈₀/J₆₈₀ KC1 lies within a narrow range, with a scattering within the range of acceptable experimental error. For this reason, the arithmetical mean of 23 experiments was computed for each relative dose of MB (0.1, 0.5, 1.0, 5, 10, and 30% of E), and a graph was constructed for the effectiveness of K-ions on a clay mineral dyed with MB, as a function of the relative dose of the dye (see Figure 2).

3. Qualitatively similar changes in spectrum curves of the MB dye on different clay minerals (montmorillonite, kaolinite, hydromica) have been observed when the suspension has been heated to higher temperatures or in the course of time (see Figures 3 and 4).

In the double adsorption hypothesis, this color effect in heating was regarded as characteristic solely of montmorillonite clays and was interpreted as a desorption of the pigment in the heating of a colored montmorillonite suspension, and as its adsorption in cooling and under other conditions, on the crystal micro-aggregates of minerals.

4. There is no desorption of pigment MB in the heating of colored montmorillonite water suspensions.

5. A relative growth of the short-wave 570 m μ maximum for the MB dye on different clay minerals (particularly of the montmorillonite group) has been observed in boiling water suspensions, prior to the introduction of the

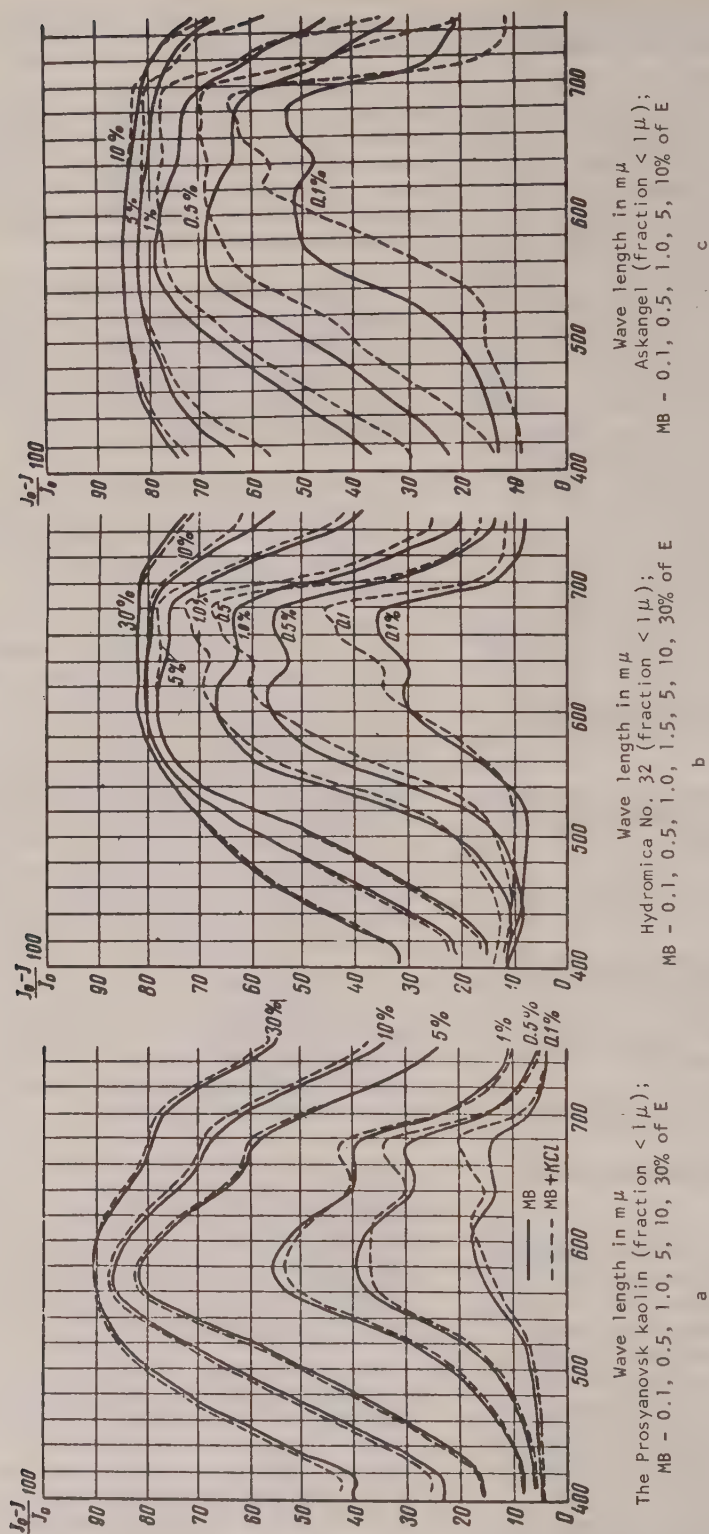


FIGURE 1. Spectrum curves of light absorption by MB adsorbed in a water medium with:

a - kaolinite (fraction $< 1 \mu$, Prosyanyovsk kaolinite); b - hydromica (fraction $< 1 \mu$ top loam No. 32); c - montmorillonite (fraction $< 1 \mu$, askangel).
Relative doses of MB: 0.1, 0.5, 1.0, 5, 10, 30% of E.

Table 1

Effectiveness of KCl on MB-colored Clay Minerals, Depending on the Saturation of the Latter by the Pigment (in % of Total Adsorptive Capacity)

Nos.	Clay Minerals (fraction < 1 μ)	Ratio J_{680}/J_{680KCl}					
		Amount of MB in % of total adsorptive capacity E					
		0.1	0.5	1.0	5.0	10.0	30.0
1	Fuller's earth	0.667	0.830	0.896	0.977	0.989	1.0
2	Gumbo clay	0.812	0.882	0.911	0.955	0.984	1.0
3	Oglaulin						
	montmorillonite	0.660	—	0.880	0.968	0.984	1.0
4	Hydromica, sample 32	0.860	0.854	0.902	0.987	0.993	1.0*
5	Kinel clay						
	(montmorillonite + hydromica)	0.829	0.953	0.966	0.993	1.0	1.0*
6.	Prosyansovsk kaolin	0.724	0.884	0.945	0.987	1.0	1.0**
7	Kinel clay (montmorillonite + kaolinite)	0.782	0.801	0.860	0.987	1.0	1.0**
8	Hydromica, sample 32	0.787	0.826	0.877	0.976	0.994	1.0
9	Latinsk clay						
	(kaolinite)	0.693	0.764	0.787	0.932	0.960	1.0
10	Lvov bentonite	0.843	0.850	0.867	0.958	0.982	1.0
11	" "	0.857	0.912	0.912	0.954	0.993	1.0***
12	Askangel	0.825	0.908	0.941	0.977	0.987	1.0
13	Hydromica + montmorillonite (No. 3)	0.882	0.927	0.940	0.976	0.995	1.0
14	Hydromica + kaolinite + montmorillonite + No. 2 (Poland)	0.798	0.906	0.937	0.979	1.0	1.0
15	" " No. 3 "	0.781	0.786	0.848	0.962	1.0	1.0
16	" " No. 1 "	0.860	0.892	0.907	0.973	1.0	1.0
17	" " No. 5 "	0.865	0.882	0.908	0.980	1.0	1.0
18	Chlorite	0.860	0.850	0.930	0.988	0.991	1.0
19	Glukhovetsk kaolin	0.897	0.949	0.960	0.988	0.995	1.0
20	Allevardite	—	0.722	—	—	—	1.0
21	Kyshtym kaolin	0.864	0.916	0.953	0.972	1.0	1.0
22	Monothermite	0.693	0.623	0.663	0.914	0.969	1.0
23	Karaganda clay (kaolinite + hydromica)	0.728	0.783	0.856	0.987	0.991	1.0
Arithmetical average		0.798	0.850	0.898	0.973	0.991	1.0

*Fraction isolated with HCl

**Without HCl treatment

***With addition of NH_4OH

ye; i. e., at the destruction of layer-structured crystal microaggregates of clay minerals. This fact is in contradiction to the N. M. Vedeneyeva hypothesis, according to which a destruction of such microaggregates should create conditions less favorable for a double adsorption of the MB pigment. Figure 8 presents changes in the MB spectrum curve for the Oglaulin bentonite, observed during intensified disaggregation in a three-minute boiling of the suspension, prior to introducing MB. A change from blue (before boiling) to violet (after boiling) was observed. Similar data for other minerals are presented in Table 2. The change in color is due to that in the J_{570}/J_{680} ratio. A rise in the latter determines the absorption intensity growth in the short-wave segment of the spectrum.

6. A comparison of spectrum lines for MB

adsorbed on different layer-structure minerals in water medium reveals the absence of hypsochrome and bathochrome effects. The color differences observed are determined by hypsochrome and hyperchrome effects; i. e., by a

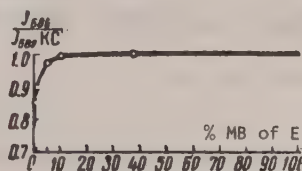


FIGURE 2. Effectiveness of K-ions on layer-structure clay minerals stained with MB, depending on the relative dose of the dye.

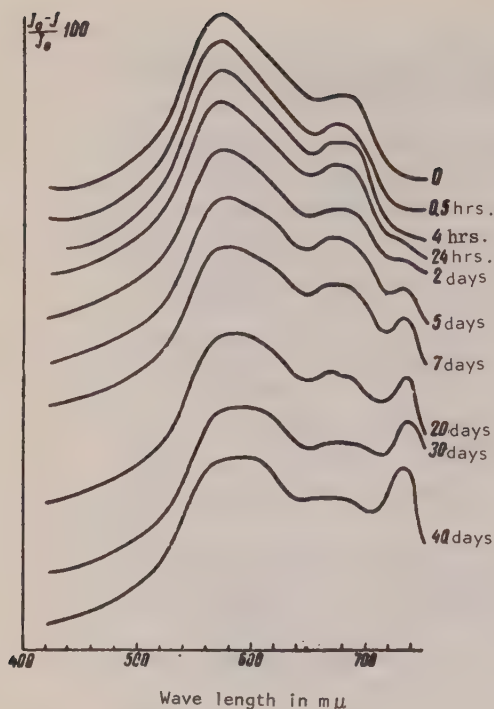


FIGURE 3. Time changes in the absorption spectrum curve for MB on montmorillonite (fraction $< 1 \mu$, Oglanlino bentonite).

weakening and intensification of adsorption with fixed maxima relative to the axis of the wave lengths. The apparent slight shift of indistinct maxima within $\pm 5 \text{ m}\mu$ is explained by a superposition of as many as six wide waves in a comparatively small segment. The divergence between the similar distinct maxima in our experiment was up to $\pm 5 \text{ m}\mu$. The same is true for parallel experiments on the same subject. The "double adsorption" hypothesis calls for a migration of maxima along the wave-length axis; i. e., for hypsochrome and bathochrome effects.

Our experimental data are explained below by surface chemical reactions between high-dispersion aluminosilicates and principal organic compounds possessing an ionic structure.

Theoretical Premises for an Interpretation of Experimental Data

The concept of surface compounds and surface reactions was clearly formulated first by N. A. Shilov [24], in 1930, in his work on the adsorption of electrolytes by coal. L. K. Lepin' [14] has established the place of surface chemical reactions among various adsorption phenomena. Surface chemical reactions are intermediate between two types of sorption phenomena: molecular or physical adsorption and chemical sorption. As in physical adsorption, surface chemical reactions do not form any new substances; but in chemical sorption, they lead

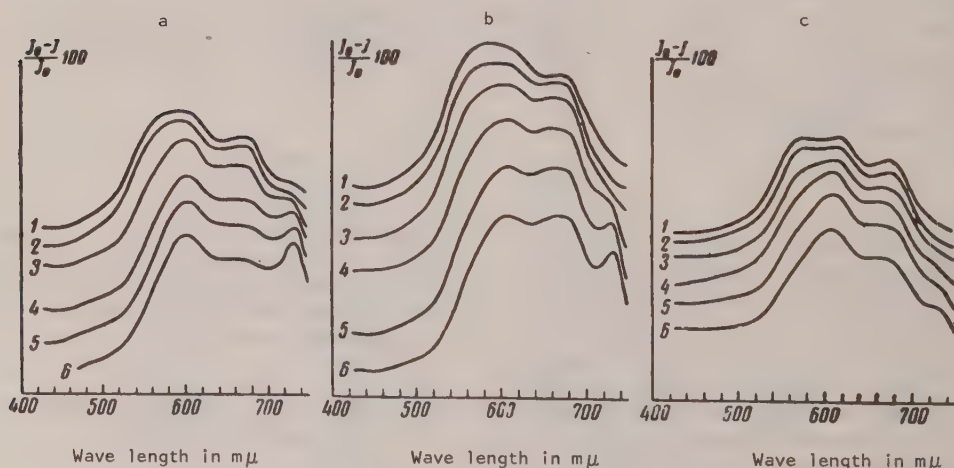


FIGURE 4. Changes in the absorption spectrum curves for MB on:

a - montmorillonite (fraction $< 1 \mu$, Oglanlino bentonite; b - hydro-mica (fraction $< 1 \mu$, top loam No. 32); c - kaolinite (fraction $< 1 \mu$, Kyshtym kaolin) - taking place in the heating of colored water suspensions: 1 - 20°C for 20 min; 2 - 40°C for 20 min; 3 - 60°C for 20 min; 4 - 80°C for 20 min; 5 - 100°C for 20 min; 6 - 100°C for 2 hrs.

Table 2

Effect of the Degree of Crowding of Clay Particles in Water Suspension
on Coloring by the MB Dye

Clays	Preparation of the suspension	J_{570}/J_{680}
Gumbo clay	Dry powder shaken in water for 10 min	1.49
	Same + suspension boiled for 3 min	1.69
Oglaulin bentonite	Dry powder shaken in water for 10 min	1.09
	Same + suspension boiled for 3 min	2.05
Na-montmorillonite (fraction 1 μ , Oglaulin bentonite) saturated with Na^+	Dry powder shaken in water for 10 min	1.15
	Same + suspension boiled for 3 min	1.43
Glukhovetsk kaolin	Dry powder shaken in water for 10 min	1.42
	Same + suspension boiled for 3 min	1.62

to chemical changes in surface atoms of a solid and in molecules of the surrounding medium. A true chemical bond emerges in such instances. The solid body atoms in a surface chemical reaction do not lose their bond with other atoms constituting the spatial lattice of the solid. Nevertheless, because of their chemical bond, surface compounds possess the chemical properties of common volume compounds, such as bases, acids, salts, etc. The formation of surface compounds proceeds at a measurable rate dependent on temperature.

The lack of uniformity in the mineral composition of clays as well as in the crystallochemistry and crystallography of their surfaces is expressed in the difference of the adsorption energy of active centers of the adsorbent. It is therefore possible that several types of salt-like surface compounds are formed as a result of surface chemical reactions between clay minerals and the MB dye: those on edges, angles, cleavage planes; i. e., in places of broken bonds, and on faces. The formation of surface compounds on faces is possible only when those faces are active with relation to the surrounding pigment molecules. In layer-structure aluminosilicates this is true for the principal pigment only when they carry a negative charge or have hydroxyl groups in the surface layer, tending to form hydrogen bonds with electrophilic dye groups [9, 25]; i. e., with groups of atoms attracting the electrons.³

Negative charges are present on basal faces of clay minerals whose crystalline lattices carry isomorphic replacements of metal atoms (montmorillonite, hydromica, chlorite, etc.). Certain layer-structure minerals such as talc and pyrophyllite have no active faces because their structures have no such isomorphic replacements; the latter are also missing on faces of surface hydroxyl groups. In kaolinite group minerals, basal faces presented by surface hydroxyl groups may be active with relation to MB. Basal faces represented by oxygen atoms cannot be appreciably active because isomorphic replacements of metal atoms are not typical of these minerals.

Interaction of the dye with clay minerals should lead to changes in the organic compound molecule, which either reduce or increase the constant (i. e., independent of the action of light) shift of π -electrons in a system of combined double bonds of the pigment;⁴ i. e., those electrons determining the double bond and belonging to the molecule as a whole. In its turn, this should be accompanied by an absorption shift to the short-wave (hypsochrome effect) or the long-wave (bathochrome) segment of the spectrum. It is known that the emergence of a chemical bond between the surface layer of the adsorbent and dye atoms should be accompanied by a bathochrome effect [19], while the formation of various associates of the dye and water

³The most reliable structural studies of MB are those of G.S. Zhdanov and his collaborators [9].

⁴Double bonds are considered combined when they alternate with single ones. In benzene chains, the bonds are of this type.

is accompanied, according to data extant [15, 22], by a hypsochrome effect.

Experimental Data

In most solvents, particularly in water, the pigment ions, beginning with concentrations of the order of 10^{-3} mol/liter, are present as dimers or double ions. The maximum absorption of a dimer ion is shifted by 1 to 50 m μ toward the short wave side, with relation to the maximum of a monomer ion. A water solution of the MG dye absorbs light in the visible part of the spectrum with maxima of 600 and 656 m μ [19]. It is believed [12, 19] that the first is typical for the dye dimers and the second for monomer molecules. The intensity ratio for these maxima depends on temperature and concentration of the solution.

As a result of spectrophotometry, by this method the water suspensions of various high dispersion aluminosilicates colored with MB, it became possible to subdivide all of the obtained surface compounds of MB into two groups. The first group is formed only on layer-structure minerals with active basal faces. Spectrophotometrically, they are characterized by the 570 and 730 m μ absorption maxima which are mutually related and change in a certain way depending on adsorption conditions. The second group is formed on the above-described minerals as well as on layer-structure minerals with inactive basal faces (talc, pyrophyllite), on chain-structure minerals (sepio-lite), and on amorphous substances (silicagel). Spectrophotometrically, they are characterized by the 620, 640, 660, and 680 m μ absorption maxima. Thus, a preliminary assumption was made concerning specific surface compounds of MB dye on active basal faces of layered aluminosilicates. Accordingly, all adsorbents studied also were divided into two groups. Those adsorbents which, in reaction with a water solution of MB, produced all of the adsorption maxima listed above in spectrophotometrically studied colored water suspensions are assigned to minerals of the first group. All layer-structure clay minerals (montmorillonite, kaolinite, hydromica), also micas, chlorites and gibbsite are included in this group. Adsorbents which, under the same experimental conditions, produced in reaction with MB the absorption maxima 620, 640, 660 and 680 m μ are assigned to minerals of the second group. Included in this group were layer-structure minerals with inactive basal faces (talc, pyrophyllite), non-layered minerals, and amorphous substances (silicagel, etc.). An overlap of a considerable number of wide absorption bands in a comparatively narrow spectrum segment (570 to 750 m μ) hampers the qualitative and quantitative analyses of the total spectrum curve. The degree of differentiation of the latter into its component maxima depend to a

great extent on the amount of dye and on the sorptive capacity of the adsorbent. With a relative increase in the dye dose, all component absorption maxima are increased absolutely; this brings about a greater overlapping effect which leads eventually to, and "smoothing-down" of, the curve. For example, in reaction of MB with water suspension of clay minerals where the dye dose rises above 10% of E (sorptive capacity), the entire orange-yellow-green part of the spectrum is absorbed more than the light-blue, blue, and red. Consequently, all such suspensions take on similar blue-violet hues, while the spectrum absorption curves show a single common broad maximum of considerable intensity decreasing gradually to a small extent, toward the red and violet; they are also characterized by the total absence of differentiation into their component maxima.

By analyzing a great number of spectrum curves for MB adsorbed in various amounts by different clay and non-clay minerals, we succeeded in determining the position of component maxima of these curves in the 550 to 750 m μ sector. Figure 5 presents diagrammatically the resolution (and addition) of total spectrum absorption curves for MB, on minerals of the first and second group, also of the absorption curve for a water solution of the dye. The most difficult to identify on spectrum curves of adsorbed MG is the 640 m μ maximum which appears to have the lowest intensity and is located in the middle where the overlapping effect of adjacent adsorption bands is particularly strong. It was found, however, on minerals of both the first (kaolinite) and the second (talc, pyrophyllite) groups. The other maxima are more distinct.

Changes in Surface Compounds of MB with High-Dispersion Aluminosilicates in Water Suspension, Dependent on Time and Temperature

The MB-colored water suspensions of the second-group minerals do not appreciably change their color, during long storage (30 to 60 days) in darkness, nor in a boiling for several hours.

The MB-colored water suspensions of the first group minerals do change their color when stored in darkness and when heated. It turned out that the temperature-induced changes are the same as those brought about by time but they proceed at a considerably faster rate. Figure 3 presents a series of spectrum absorption curves for a colored water suspension of montmorillonite, obtained by time spectrophotometry at 20°C. Similar curves were obtained for hydromica and kaolinite suspensions. The graph shows a gradual appearance and growth of the 730 m μ absorption maximum and a relative decrease of the 570 m μ maximum.

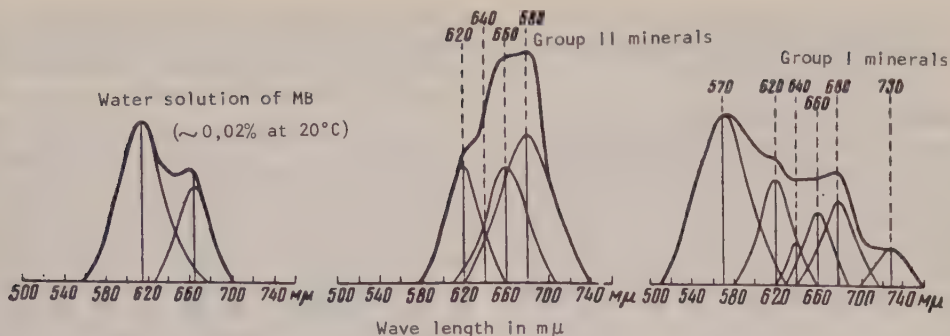


FIGURE 5. Diagrams of absorption spectrum curves for MB, in water solution and in the adsorbed state on high dispersion aluminosilicates in a water medium.

Measurements of the other maxima and their ratios show that their relative sizes remain almost unchanged.

In order to determine the effect of temperature on the change in color of the first mineral group suspensions (montmorillonite, hydro-mica, kaolinite), the latter were successively kept for 20 min, at temperatures of 20, 40, 60, 80, and 100°C; in addition, they were boiled for two hours. The corresponding series of spectrum curves for various minerals are presented in Figure 4. They show that temperature-induced changes are similar to those taking place in the course of time. The difference is that these suspensions have to be stored for one month, at 20°C in order to produce changes which occur after two minutes of boiling.

Changes in Surface Compounds of MB With High-Dispersion Aluminosilicates In the Process of Dehydration and Hydration of Dyed Samples

No appreciable color changes take place in the drying of colored water suspensions of the second group of minerals; what changes there are consist of an abrupt drop in the intensity of the 570 and 730 mμ absorption maxima, in drying the suspensions of the first group of minerals. The relative value of these maxima depends on hygroscopic moisture content of the colored mineral powder. When a colored powder is hydrated gradually and at a higher relative humidity of air, the 570 and 730 mμ maxima, too, increase gradually. The study has revealed a linear relationship between J_{570}/J_{680} and the moisture content of the dyed clay sample. For instance, it was $J_{570}/J_{680} = 0.0075 W + 0.555$ for the Oglanlin bentonite colored with MB in the amount of 5% of E, where W is the moisture content of the dyed sample, in %, as determined in drying at 105°C.

No color changes have been observed in

moistening the colored powder of the first group of minerals with nonpolar organic liquids.

Effect of the Pigment Dose on the Results of Coloring Water Suspensions of Aluminosilicates

The color of water suspensions of high-dispersion aluminosilicates dyed with MB depends to a great extent on the amount of dye in reaction with the minerals. This effect is particularly conspicuous for the first group minerals which include most clay minerals. This is explained on the whole by two facts: 1) the lack of uniformity in the adsorption energy for active centers of adsorbing minerals; and 2) by the physical conditions of overlapping of a considerable number of wide absorption bands within a narrow 550 to 570 mμ spectrum range.

It is believed that the adsorption potential of active centers of clay minerals, located on crystal edges, angles, and cleavage planes, is higher than in active centers on crystal faces. The graph in Figure 1-a shows clearly that, at small dye doses, absorption in the long-wave interval is the more intensive (maxima 620, 640, 660, and 680 mμ); this is a characteristic of surface compounds of the dye with active centers of crystal edges, angles, and cleavage planes. A water suspension of various clay minerals is colored here in greenish hues, changing to light-blue and blue with a higher dose of the dye.

A further increase in the dose brings about an absorption maximum in the short-wave interval (570 mμ); it characterizes the surface compounds of the dye with the active basal faces of clay minerals. The suspensions take on violet hues. The color change dependent on the dye dose is less conspicuous in the second group of minerals, and is explained on the whole by the second of the two tendencies.

Effect of Changes in the Degree of Dispersion and Grouping of Minerals on the Coloring of Water Suspensions

According to spectrophotometric data, a change in the degree of dispersion and grouping of minerals of the second group has no effect on the adsorption of MB in water suspensions but merely leads to a change in the sorptive capacity of adsorbents. A change in the same factors for the first group minerals is accompanied by considerable changes in spectrum curves of the dye adsorbed.

The adsorption of MB by finely crystalline well-defined minerals of the first group is accompanied by a comparative predominance of surface compounds on basal faces. Spectrophotometrically, this is reflected in a predominance of absorption in the short-wave interval, with a maximum of 570 mμ. The adsorption of MB on high dispersion minerals of the first group obtained by mechanical crushing of their coarser crystalline forms is accompanied by an increase in the number of surface compounds on bond breaks of deformed crystals. Spectrophotometrically, this is expressed in a predominance of long-wave absorptions (620, 640, 660, 680 mμ). We have illustrated this phenomenon for muscovite (see Figure 6). Muscovite tablets from the Mamsk deposits were crushed in an eccentric mill, for 10 minutes the first time, and to an amorphous state in the second. The sorptive capacity of muscovite was changed from 8 to 13 mgm/equiv to 100 gm. Changes in the absorption intensity ratios for 570 and 680 mμ were determined from a graph (Figure 6) constructed from spectrophotometric results. This graph shows that

surface compounds of MB with amorphous muscovite are characterized by a relative shift to absorption toward the long-wave segment (a reduction of ratio J₅₇₀/J₆₈₀).

The degree of grouping of mineral particles of group one (i. e., most clay minerals) has quite a strong effect on the spectrum curve of the dye adsorbed. In most instances, aggregation of layered minerals present tightly packed tablets of primary particles. Their disaggregation in a water suspension, by various methods (boiling, supersonic action, etc.) increases the amount of crystal tablets and consequently the over-all surface of basal faces. This leads to the formation of a comparatively large number of surface compounds of the dye with active basal faces, wherein the MB colored suspension takes on violet hues. Table 2 shows the absorption intensity ratio for maxima 570 and 680 mμ for clays of a different degree of grouping in a water suspension.

This table shows that the color change in suspensions is quite large. The graph in Figure 7 presents absorption curves for colored Oglanlin bentonitic clay, at different degrees of grouping for the particles in a water suspension. This graph shows that a disaggregation of the adsorbent particles is accompanied by an increase in the intensity of the short-wave (570 mμ) absorption maximum.

Effect of Iron Hydroxide Films at the Surface of Clay Minerals on the Coloring of Water Suspensions with MG Dye

A positive sol of iron hydroxide was obtained by the Grimm method. A 1.3% iron hydroxide

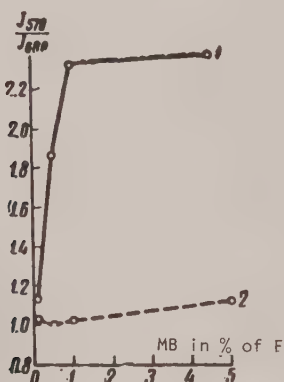


FIGURE 6. The effect of higher dispersion of muscovite on J₅₇₀/J₆₈₀, depending on dose of MB dye

1 - muscovite ground up in eccentric mill, for 10 minutes; 2 - muscovite ground to amorphous state.

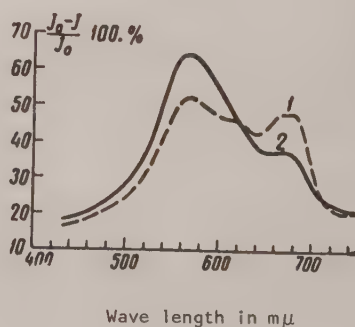


FIGURE 7. Effect of the degree of packing in Oglanlin bentonite, in water suspension, on the nature of its coloring by MB dye

1 - suspension prepared by shaking bentonite powder with water, for 10 minutes; 2 - same with an additional boiling for 3 minutes.

ol (assuming its concentration in suspension minerals at about 13%) was added to their 2% water suspensions. After a two day settling, the transparent solution was decanted, and the sediment was dried at 60°C and then ground to powder. Spectrophotometric results have shown that ferruginous minerals form a relatively smaller number of surface compounds with the dye, typical of basal faces. Likewise, spectrophotometric results for natural clays colored with MB and carrying a considerable amount of amorphous iron hydroxide (up to 11% in fraction $<1\mu$) have shown a considerably smaller number of these surface compounds.

This can be explained partly by the high water resistance of ferruginous crystals of clay minerals and partly by a lowering of activity in basal faces because of the adsorption of $\text{Fe}(\text{OH})_3$ from the positive sol.

Specific Features of Coloring Polymineral Mixtures

Different crystallographic elements of clay mineral crystals are known to possess different adsorption energies. Because of that, in dyeing polymineral mixtures, such as all natural clays, it is most probable that the dye is adsorbed by crystals of all minerals but in different proportions, depending on the adsorption energy of active mineral centers. To demonstrate that experimentally, binary mixtures were prepared of montmorillonite (fraction $<1\mu$, askangel) with 10% sepiolite; with 50% talc; and with 50% hydromica. Spectrophotometric data were used in constructing the graph of changes in J_{570}/J_{680} dependent on the dye dose (see Figure 8). A correlation of such binary graphs with the corresponding graphs for individual components

leads to the conclusion that the two minerals are colored simultaneously with different doses of the dye. This is because in the dyeing of a single mineral, for example montmorillonite alone, the dye dose would have increased relatively, in proportion to the amount of the mineral; this, in turn, would have led to a higher J_{570}/J_{680} ratio, which is not the case. On the other hand, if talc and sepiolite alone did take on the dye, the color effect of the introduction of K-ion in suspension would have been missing at small dye doses, again contrary to the presence of that effect in all instances. It follows then that an identification of mineral mixtures by the staining method and spectrophotometric analysis cannot be based on a classification of minerals by their adsorption energy.

Effect of Ions K^+ , Rb^+ , Cs^+ , and NH_4^+ on MB Colored Water Suspensions of Aluminosilicates

N. Ye. Vedeneyeva has noted that large ions of alkali metals (K^+ , Rb^+ , and Cs^+) as well as ammonium ions, upon being introduced in a colored water suspension of certain clay minerals (mostly montmorillonitic, according to N. Ye. Vedeneyeva), bring forth a certain color effect expressed in an intensification of long-wave absorption and in a weakening of the short-wave absorption. On the basis of this double adsorption of MB dye, N. Ye. Vedeneyeva believed it to be unstable in montmorillonite minerals only; the effect of K-ions she reduced to the creation of conditions unfavorable for the appearance of "pseudocrystalline scales of montmorillonite" on which this type of adsorption is possible. According to N. Ye. Vedeneyeva [4], double adsorption is always possible on larger kaolinite particles with better

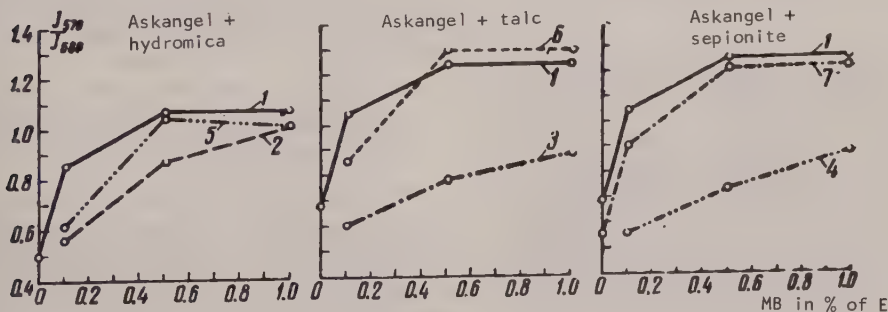


FIGURE 8. Ratio of the absorption intensity, J_{570}/J_{680} as a function of the dye dose in spectrophotometry of MB colored water suspensions of synthetic mixtures of minerals

1 - askangel (fraction $<1\mu$); 2) hydromica No. 32 (fraction $<1\mu$); 3 - talc; 4 - sepiolite; 5 - askangel +50% hydromica; 6 - askangel +50% talc; 7 - askangel +10% sepiolite.

crystalline forms. This has not been corroborated by our study.

The effect of K-ions is reduced with a higher dye saturation of all minerals of the first group (see Figures 1 and 2), because of a probable reduction in the number of exchange places accessible for these ions. Minerals of the second group show no color effect whatsoever, in the action of these cations. At an MB saturation of over 10% of E, for various clay minerals, spectrophotometric analysis reveals practically no color changes at the introduction of K^+ , Rb^+ , Cs^+ , and NH_4^+ ions (Figures 1 and 2; Table 1).

Although this phenomenon is quite complex, still the spectrophotometric data lead to certain assumptions. The introduction of K-ions in a colored suspension, is accompanied by a relative lowering in the intensity of absorption maxima 570 and 730 $m\mu$; i.e., by the same situation prevailing in dehydration of the colored sample.

According to various authors, when layer-structure clay minerals are saturated with K^+ , Rb^+ , Cs^+ , and NH_4^+ ions, the latter are arranged mostly on basal faces. The hydration of clays is affected by the cations adsorbed. According to O. Ya. Samoylov's data [18], all those cations possess a "negative" hydration; i.e., they intensify the translation movement of water molecules nearest to an ion. It is known, for example, that K-replacement clays are definitely the least hydrophilic [10, 17]. Spectrophotometric studies of the dehydration-hydration process in dyed minerals of the first group suggest that color effects of the introduction of those cations in a MB dyed water suspension of layer-structure clay minerals are related to a change in the hydration state of basal faces of these mineral crystals and possibly to dehydration of salt-like surface organic-mineral compounds originating in the adsorption of MB.

The Adsorption Hypothesis for the Basic Organic MB dye on High-Dispersion Silicates

Surface salt-like compounds of two types are formed as a result of the reaction of thiazine dye MB with high-dispersion aluminosilicates in a water medium. Salt-like hydrate compounds on active basal faces of layer-structure minerals are characterized by light absorption in the short-wave interval, with a maximum at 570 $m\mu$. The hypsochrome effect, compared with a water solution of MB, is not in contradiction with the data on hand [15, 22] which suggest that this very effect is brought about by an association of MB and water. The formation of these surface hydrate compounds proceeds at a measurable rate dependent on temperature. The gradual strengthening of the chemical bond between the dye and the adsorbent is marked

spectrophotometrically by a bathochrome effect while the intensity of the 470 $m\mu$ maximum is relatively lowered and that of the 730 $m\mu$ is raised. A rise in temperature of the colored suspension considerably speeds up this process. Drying the colored water suspension leads to a reversible dehydration of surface hydrate salt-like compounds. Their dehydration is marked spectrophotometrically, by a reduction and complete disappearance (in drying at 100°C) of the 570 and 730 $m\mu$ absorption maxima. It appears that salt-like surface compounds are formed at the same time, without the constitutional water. According to spectrophotometric data, the color effect of K-ions introduced in the MB colored suspension, and consequently in the mineral exchange complex, is similar to a partial dehydration of surface hydrate compounds of MB dye; i.e., the intensity of the 570 and 730 $m\mu$ adsorption maxima is reduced with a reduction of the number of places accessible for K-ions, at the adsorbent surface, as the dye dosage is increased. This explanation of the color effect of K-ions is not in contradiction to what we know of the adsorption of these ions by clay minerals, and of their properties with relation to water molecules [7, 16, 17, 18].

The second type of surface chemical compounds of MB is formed in water suspensions of layer-structure aluminosilicates, at places of broken bonds, on crystal edges, angles, and cleavage planes, as well as on non-layered minerals (chain-like, etc.) and on amorphous adsorbents. Spectrophotometrically, this type is characterized by absorption in the visible spectrum, with maxima of 620, 640, 660, and 680 $m\mu$. Such compounds are stable in time and do not change at the higher temperatures of colored suspensions.

Thus it becomes possible to divide the silicate minerals into two groups, on the basis of their surface compounds with MB. All layer-structure aluminosilicates with active basal faces belong to the same group.

The Staining Method as a Means of Mineral Analysis of Clay Rocks

The great achievement of N. Ye. Vedeneyeva is her use of spectrophotometric analysis in the study of adsorption of thiazine dyes on clay minerals and the establishment of principal regularities in their staining. We regard our own study as a direct continuation of N. Ye. Vedeneyeva's work on the adsorption of basic dyes by various minerals. Our deviation from her methodology does not mean we have rejected it, but that ours is rather a more comprehensive study of this complex process.

Irrespective of any theoretical interpretation of spectrophotometric data on the adsorption of

the MB dye by various high-dispersion aluminosilicates, practical conclusions can be stated about this method as a means of mineral analysis of clays.

First, the nature of this main diagnostic criterion for the staining methods, the color effect of K-ions, can be described as follows. As illustrated in the graph of Figure 2, the sorption capacity of an adsorbent can be determined by measuring, from the spectrophotometric data, the relative quantity $J_{680}/J_{680}KCl$, by a rigorously quantitative experimental procedure (the amount of adsorbent and dye). Consequently, the color effect of K-ions is an indirect qualitative characteristic of that portion of the total sorption capacity of clay minerals determined by active adsorption centers of the basal faces of layer-structure clay minerals. According to our own data for such minerals, this portion constitutes about 90% of the total adsorption capacity.

Additional experiments corroborate our conclusions on the nature of the color effect of K-ions. Figure 9 is a comparative description of a lithologic column (test No. 1005 in the Kursk Magnetic Anomaly area; data by the MOSGEO), as obtained by the Vedeneyeva-Vikulova MB staining method with visual registration, and a quantitative determination of the sorption capacity of those rocks (hygroscopic moisture content and sorption capacity for MB, converted for the adsorption-active rock fraction). The staining and the determination of hygroscopic moisture content was done by F.I. Krasnovskaya.

Figure 9 shows clearly that all three criteria (the color effect of K-ions in relative units; the hygroscopic moisture content in %; and the sorption capacity in mgm/equiv. for 100 gm) are directly related to each other and characterize the sorption properties of clay minerals. In some monomineral clays the hygroscopic moisture content and sorption capacity, expressed in absolute units, may serve as indirect criteria in determining the mineral composition of clays. The color effect of K-ions, expressed in relative units of an arbitrary ten-point color scale, is only a qualitative characteristic of the sorption capacity of clay minerals; this creates additional difficulties in the use of that effect in mineral identification.

It now becomes clear why the staining method will not identify kaolinite and hydromica in a mixture with montmorillonite, even when their content in a binary system is 80 to 90%. A small addition of montmorillonite with its high sorption capacity of the mixture to where, at the "standard staining" (MB_{st}) recommended in this method, the amount of MB is considerably less than 10% of E (see Figure 2); i. e., to where the K-ion effect appears.

Second, it may be assumed that the total

color characteristic of MB dyed water suspensions of clay minerals is a complex function of many variables, only a part of which has been studied. It has been established that at different combinations of variables (not all of them subject to standardization under our experimental conditions) the same clay mineral can acquire almost all the colors of M. F. Vikulova's ten-point scale. Analyzed with the SF-2M spectrophotometer, the MB dye adsorbed by various layer-structure clay minerals is characterized qualitatively by the same type of absorption curves (with maxima on the long-wave axis). All this leads to the conclusion that the total color characteristic of MB adsorbed by clay minerals cannot be utilized for either quantitative or qualitative mineral analysis of clays.

Third, all actual data suggest that the spectrographic study of adsorption of basic dyes by clays is suitable in solving specific problems related to the kinetics of changes in surface properties of clays. For instance, the above-mentioned linear relationship between the relative quantity J_{570}/J_{680} (for a given relative coloring of a clay) and its moisture content can be used in registering the moisture content changes in clays, in the course of an experiment. The relationship between the micro-aggregate composition of a clay (Figure 7) and the color effects in MB dyed water suspensions can serve as a criterion in estimating the water resistance of clay microstructure. The color effect, illustrated in Figure 6 for MB on a layer-structure aluminosilicate, as a function of its mechanical crushing, can serve as a criterion for the degree of such crushing in layer-structure minerals, for some specific purpose. Chances are that spectrophotometric analysis can be used in other fields of practical application for color reactions of a dye in clays.

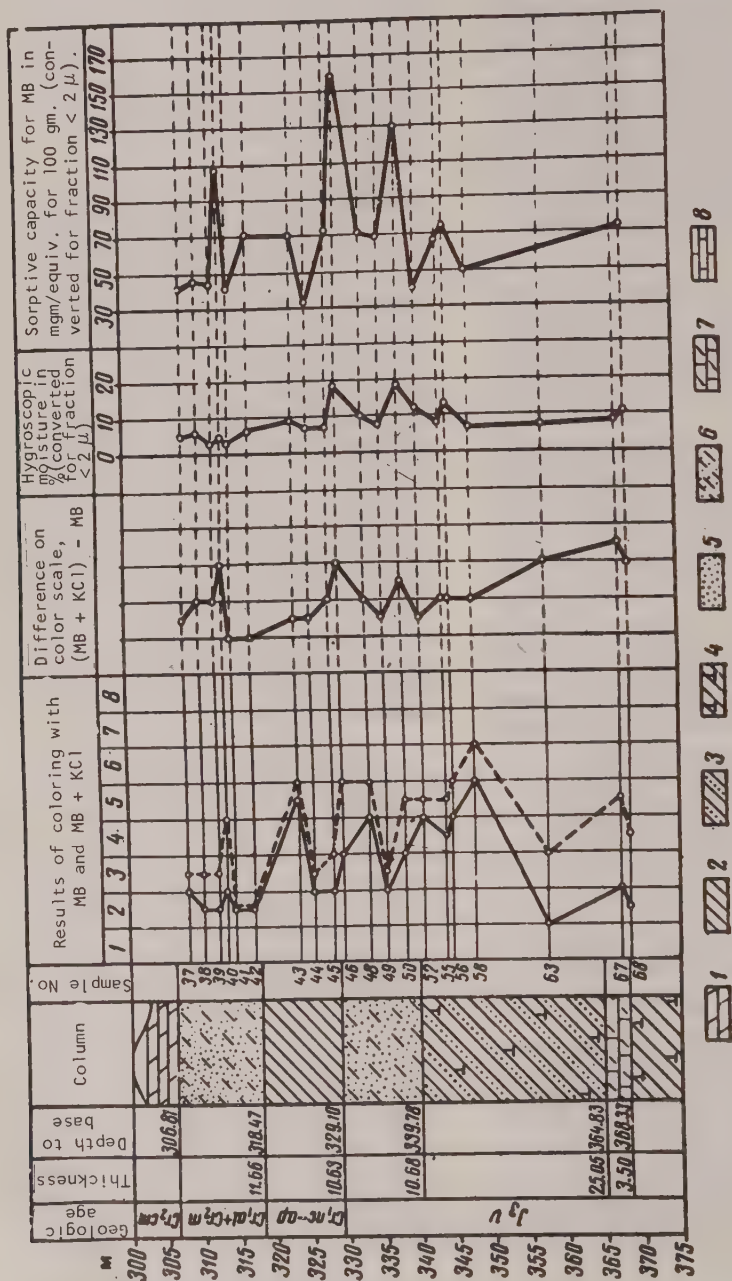
Fourth, the experimental data cited above, and the consequent conclusions arrived, make the relative color effect of K-ions in using the MB dye, the "rational core" of N. Ye. Vedeneyev's method. However, it should be agreed upon that this effect is characteristic not of the structural properties of clay minerals but rather of their sorption capacity which is taken to be, in this study, an indirect criterion in estimating the mineral composition of clays. It should be kept in mind, however, that this criterion is not quite reliable because the sorption capacities of different clay minerals vary in a fairly broad range [7]. On the other hand, this criterion is unquestionably valid in differentiating clay sequences into natural groups for a further mineral analysis (X-ray, thermal, etc.). Future efforts should be directed toward developing simpler and more rapid quantitative methods of determining the adsorption capacity of clays.

One such method, fairly simple and accurate, calls for the use of MB dye [2, 13, 23]. If the

Vedeneyeva-Vikulova method is used, it is more justifiable, both theoretically and practically (see Figure 9), to construct the lithologic columns by the relative color effect of K-ions, expressed graphically by a single line from points obtained by subtracting the total MB color characteristic from that of MB + KC1 (on the ten-point scale).

As of now, M. F. Vikulova believes that the

staining method is of prime importance among other mineral analysis methods for clays. In her 1960 work [6] she states that "the petrographic, X-ray, and staining methods are paramount in the study of clay; all others methods are merely supplementary". She regards the staining method as an indirect one in qualitative mineral identification of clays. Such a view contradicts M. F. Vikulova's own data [6] and has not been corroborated by our studies



Any method of a qualitative mineral analysis of clays should satisfy this basic requirement: a qualitative determination of clearly predominant minerals in a natural polymineral clay mixture. Judging from M. F. Vikulova's graphic material on the probability of identifying clay minerals in mixtures, by various methods, the staining method, taken by itself, does not fulfill this requirement. For instance, in mixtures of 80 to 90% kaolinite and 10 to 20% montmorillonite or 80 to 90% hydromica and 10 to 20% montmorillonite, neither one of the three will be detected by the staining method, so that the clay mineral type will be determined incorrectly.

Our new experimental data considerably restrict the application of this method of clay analysis. However, it demonstrates the prospective suitability of N. Ye. Vedeneyeva's spectrophotometric analysis as a means of applying the adsorption analysis of clays in the study of the kinetics of change in their surface properties, as well as in some problems in the ion exchange, and in other current problems of clay study.

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REVIEWS AND DISCUSSIONS

CRITICAL OBSERVATIONS OF P. P. SMOLIN'S BOOK, "PRINCIPLES OF RATIONAL CLASSIFICATION OF METACARBONATE ROCKS"^{1, 2}

by

A. S. Varlakov

P. P. Smolin's classification of metacarbonate rocks is based chiefly on qualitative mineral criteria. The author uses a number of familiar terms, in definitions of his own; on the other hand, he rejects many popular terms such as ophiocalcite, forsterite marble (and silicate marbles in general); the independence of brucite marbles he regards as "petrographically unjustified".

We shall consider this classification in the order arranged by the author. We pause first on the definition of its quantitative boundaries. P. P. Smolin takes 10% as the maximum content of accessory minerals, which, in his opinion, "corresponds to the usually accepted maximum content of accessory minerals in rocks". However, the data published on this subject are at variance with P. P. Smolin's view. It is well known that authors of classifications for sedimentary carbonate and terrigenous carbonate rocks accept 5% as the limit for accessory minerals' content. Logically and methodologically, it would be correct to use the same figure for metacarbonate rocks, as well.

P. P. Smolin designates three large groups: marbles (with not over 10% silicates); calcifers (10 to 90% metasilicates); and hornfels and skarn (not over 10% carbonates). It appears then that the silicate marble group is assigned here to "calcifers".

Differentiated within each group (with the

¹Kriticheskiye zamechaniya k stat'ye P. P. Smolina "Ob osnovakh ratsional'noy klassifikatsii metamorfizovannykh karbonatnykh porod".

²Izvestiya Akad. Nauk S.S.S.R., ser. geol., no. 12, 1959.

exception of hornfels and skarn) are varieties dependent on the ratio of carbonate minerals to an arbitrary pencatite component. This latter term means that portion of rock consisting of calcite and brucite, in equimolecular relationship. It should be noted in this connection that the designation of pencatite as an independent component, along with dolomite and brucite, introduces a parameter inconsistency. In so doing, it becomes necessary to indicate the brucite content in a rock in each case; incidentally, its qualifying limit is taken to be 4% rather than 10%.

Named in the marble group are 1) essentially carbonate rocks (with not over 10% pencatite component); 2) brucite marbles and essentially brucite rocks; and 3) borate marbles.

Recognized among the essentially carbonate marbles are dolomitic varieties, with as much as 10% magnesite. And while it is recommended to introduce the silicate name in that of a marble with as much as 10% of that silicate, this rule is relaxed for magnesite and brucite. Generally speaking, by adhering to the quantitative criterion, P. P. Smolin obscures the genetic aspect of the nomenclature, the aspect that usually is the most important in marbles. Even when a marble carries less than 4% brucite, its very presence should be emphasized more than, let us say, that of 10 or 15% of muscovite, because it is indicative of certain specific features of metamorphism, besides mere crystallization. P. P. Smolin ignores that fact.

Nor can we agree with the assigning of magnesite to the marble group, for no good reason. In any event, magnesites, although carbonate rocks formally, have a place of their own. It is often quite difficult to decide whether a magnesite is a primary sedimentary formation, subsequently recrystallized, or a metasomatic one. This subject has been argued for decades. When hydrothermal magnesites are assigned to the marble category, the principal genetic criterion of the term, "marble", is lost; by the same token, a magnesite (definitely known to be primary) assigned to marbles should be called, "magnesite marble" rather than "magnesite".

P. P. Smolin divides the brucite marble subgroup into two main varieties, pencatite and predazzite, destined to replace the "petrographically unjustified" term, "brucite marble", of the current nomenclature. These terms were originally assigned, on the basis of chemical analyses, to two "minerals" — CaCO_3 , $\text{Mg}(\text{OH})_2$ and $2\text{CaCO}_3 \cdot \text{Mg}(\text{OH})_2$. It has been subsequently established microscopically that they are not minerals but rocks consisting of calcite and brucite in appropriate molecular relationship. In most natural occurrences, the brucite-calcite relationship is diverse, first because brucite may have a different primary dolomite content; second, because dedolomitization processes are often accompanied by a leaching or an addition of magnesium and partly of calcium. It can be stated that pencatites as such are almost nonexistent in nature, and the term itself is used somewhat arbitrarily. Thus, the original meaning of the terms pencatite and predazzite, as beds with a definite composition, has been completely lost. It follows that the use of such terms does not make any sense, and the procedure should be the reverse of that proposed by P. P. Smolin. In striving to preserve the original meaning of these terms, he follows a formal path and accepts the pencatite component as a basis of his subdivisions. It turns out, as a result, that the brucite content may be larger in predazzite than in pencatite; and that such varieties as "dolomitic pencatite" and "dolomitous predazzite" cannot be differentiated at all by the amount of brucite. By the way, the introduction of such definitions as "dolomitous" is hardly felicitous. The need for such wordmongering has been disproved, some time ago.³

For rocks with a brucite content greater than that possible for pencatite, the author introduces the term, "marble". As understood by P. P. Smolin, this group includes "brucite"; i. e., a rock no less than 90% brucite. How this rock happened to be included with marbles is known only to the author. The same is true for "brucite-magnesite". One cannot condone the designation of an almost monomineral brucite rock as "brucite". Any way we look at it, brucite is a mineral, so that the rock consisting almost wholly of it should be called, in the conventional nomenclature, either brucitite (in analogy with serpentinite, talcite, etc.) or else brucite rock.

P. P. Smolin is right in stating that a classification should not attempt to name all natural rocks. However, he has specially designated such varieties as "brucite-magnesite" and "cotoitic marble" as rocks occurring in rare

cases. Nevertheless, periclase marbles are missing for some reason.

Also questionable is the expediency of such terms as "protobrucite pencatite", inasmuch as the genetic nature of brucite is commonly difficult to establish, and the very name, "pencatite", implies the presence of brucite. When however, the primary nature of brucite has been established, it is more properly called "protobrucite marble". The term, "apopericlase brucite marble", is etymologically wrong because the prefix, "apo", is ordinarily added to rock names rather than to mineral names.

The author's calcifer group includes rocks of different origin, in conformity with his quantitative criterion. The term, "calcifer", was first proposed by Brognard for silicate marbles which originated by metamorphism of limestone. Later on this name was automatically assigned by some authors to silicate marbles formed out of dolomite. That very fact, let alone the etymology, should inspire a critical view of this term; but the author could not think of anything better than to make it one of his basic terms.

According to P. P. Smolin, "pure" calcifer is calcite + silicate; i. e., it carries some calcite, in conformity with the etymology of this word. Inasmuch as that author has given it a universal meaning, by applying it to all metamorphic silicate carbonates in general, he has unavoidably arrived at an erroneous designation of "dolomitic calcifer", a rock which may have no calcite.

In addition, there are specific shortcomings in subdivisions of that group: for example, there is virtually no difference between "dolomitous" and "dolomitic" calcifers. The term, "brucite-carrying calcifer", is etymologically wrong, for a brucite content up to 29%. The word, "carrying", is applied ordinarily to accessory rock minerals. Again, the difference between near-skarn calcifers and common skarns is not clear, and the term, "near-skarn", is not defined. To call, those metasomatic calcifers without either spatial or genetic relationship with skarns "near-skarn", is obvious nonsense. In the presence of such a relationship, these rocks are much more conveniently named, "skarn marbles", instead of trying to justify the term selection by referring to authorities.

It can be stated offhand that the term, "calcifer", as now understood (especially in Smolin's classification), should be abandoned. It is not justifiable, either etymologically or historically, and not even petrographically. We can get along just as well without it (as the American geologists do).

The last group in P. P. Smolin's classification of metacarbonate rocks is represented by

³Yu. M. Scheinmann. Necessary Terms and "Necessitous". Byul. Mosk. Ob. Isp. Prir., otd. geol., vol. 30 No. 2, 1955.

hornfels and skarn. It is superfluous to demonstrate that these two varieties should never be united in a single group. The author himself is conscious of his weak position and does not attempt any detailed subdivision of this group, other than magnesian and calcareous.

To summarize, it is perfectly obvious that this classification can be applied only with a laboratory study of rocks; i. e., from the results of chemical analyses, quantitative mineralogic computations, etc. That means that a field geologist will have to use some other classification; that, in turn, means confusion.

However, even after a laboratory study, the geologist will still have his difficulties. Before he can arrange his rocks on the several shelves of P. P. Smolin's nomenclature he will have to perform a number of mathematical computations (for the pencatite component; the amount of calcite, not equivalent to brucite; etc.).

It often happens that most rocks designated by Smolin can be found within a small area of metacarbonate rocks. Or, varieties of all three groups may occur in ten sections prepared from a single hand specimen. This shows up the formalistic quantitative approach of P. P. Smolin's classification, disregarding the specific geologic conditions. Such a classification is not likely to become popular.

It is natural to inquire the way of a rational classification of metacarbonate rocks. It seems to us that we do not need any new classification, since we already have one which is commonly agreed upon. It is beside the point that some students, careless in their petrographic nomenclature, allow some distortions, deviations, and even illiteracy, to creep in. As an example of such an irresponsible approach, there is the common use of the term "brucite limestone", by foreign geologists. The same is true for the notorious term, "calcifer", as understood recently by Soviet geologists.

By adhering to the current terminology and to common rules of nomenclature, a correct name, understandable to every one, will be found for any rock. That is what we should be striving for, instead of working out descriptions for each particular example, as P. P. Smolin does.

To be sure, some improvements are desirable in the nomenclature of these rocks. As we have noted before, the designation of calcifers, pencatites, and predazzites is not petrographically justifiable. To be specific, I shall attempt to systematize the general principles of a petrographic classification for metacarbonate rocks.

Following the above-mentioned considerations, the essentially carbonate marbles are best subdivided in analogy with the corresponding

sedimentary rocks (as in M. E. Noiniskiy's classification), by accepting the same quantitative limit of 5% for accessory minerals. The presence of a typical mineral, silicate or any other (brucite, periclase, magnesite, etc.), up to 5%, can be introduced when necessary in the rock name, in a way accepted for accessory minerals; i. e., by adding the suffix, "carrying". Where the primary sedimentary origin of magnesite has been demonstrated, the name, "magnesite marble", is acceptable with the same intermediate quantitative subdivisions as for dolomitic and calcitic marbles.

Rock that has undergone decarbonization should be called marble, with the newly-formed mineral as an adjective (regardless of the composition of that mineral). A carbonate mineral, too, may be introduced in the name, for petrographic considerations. The general rule is to place the minor mineral first. For example, "brucite marble" is a rock consisting of brucite and calcite, with the brucite content not exceeding that possible in dedolomitization of dolomite. With this procedure, there is no need to emphasize the quantitative content of brucite in the rock name. The rock is named for its principal typomorphic minerals, with consideration given to the petrogenetic criterion, first, and then to the quantitative. All details of composition will be brought out in the description of the rock. Such rock names in the P. P. Smolin classification, as "spinel-forsteritic brucite-carrying dolomitic calcifer", are not much of a convenience.

When the amount of minerals formed in decarbonization exceeds the original; i. e., when there is an addition or a metamorphic differentiation, the term, "marble", should be substituted by "rock" or by "schist", in the presence of schistosity. For example, a rock consisting of calcite and forsterite, with the content of the calcite over 45% (the theoretically possible limit, in a complete dedolomitization of dolomite), should be called calcite-forsteritic rather than forsteritic marble.

We believe it inexpedient to designate as "hornfels" essentially silicate rocks, regarding the latter as an extreme silicate variety of metacarbonates, first because hornfels are peculiar rocks primarily characterized by their contact metamorphic origin. In P. P. Smolin's classification, hornfels would include rocks utterly unrelated to contact metamorphism. In so doing, he in effect obliterates the difference between hornfels and crystalline schists which are the product of regional metamorphism. We cannot agree with P. P. Smolin's statement that "the historic division of silicate metamorphic rocks into massive hornblendes (largely contact rocks) and crystalline schists (regional metamorphic rocks) often becomes purely formalistic". In the final count, it is not the name that matters; if it is built upon a structural

criterion, that criterion should be interpreted first in its genetic aspect. Any specific hornfels may be schistose, and a regionally metamorphosed rock may be massive, as correctly noted by P. P. Smolin. To be sure, in many cases it is difficult to draw a line between those rocks; this, however, is a different and more general problem.

It appears then that there is nothing new in the principles of the metacarbonate rock

nomenclature of P. P. Smolin; they are elementary and familiar. A correct use of these principles, with consideration given to the specific geologic conditions, will obviate the need for a special classification.

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CHRONICLE

JOINT SESSION OF ACADEMIES OF SCIENCE OF THE U. S. S. R., AND THE ARMENIAN, GEORGIAN, AND AZERBAIDZHANIAN S. S. R.

The Joint Session, commemorating the 40th year of Soviet power and the Communist Party in Armenia, was held at Yerevan, December 25-29, 1960. Besides the leading scientists from these Academies were members of other national academies of the Soviet Union as well as representatives of the Party and academic circles of the Armenian S. S. R. participated.

The Session opened with welcoming addresses by Academician A. N. Nesmeyanov, President of the U. S. S. R., Academy of Sciences; Academician N. I. Muskhelishvili, President of the Georgian Academy of Science; and Yu. G. Imedaliyev, President of the Azerbaydzhanian Academy of Science. President V. A. Ambartsumyan, of the Armenian Academy of Sciences, reported on the achievements of Armenian scientists, in the 40 years of Soviet power. Read before the general session were numerous greetings from the scientific, party, and social organizations of the country.

The work of the session was carried out in the following sessions: Physics and astronomy (10 papers); mathematics and mechanics (13 papers); technical sciences (12 papers); geologic sciences (11 papers); chemical sciences (9 papers); biologic (15); medico-biologic (20); historical (15); economics and philosophy (8); literature and languages (9).

The following papers were read and discussed in the geologic section:

1. Academician, D. I. Shcherbakov: New data on the Antarctic.

2. Academician, A. N. Dzhevakhishvili, Georgian Academy of Sciences: Natural Landscapes of the Georgian S. S. R.

3. Academician I. G. Magakyan, Armenian Academy of Sciences: Regularities in the Distribution of Rare and Dispersed Metals in Armenian Ores.

4. G. D. Afanas'yev, Corresponding Member Academy of Sciences, U. S. S. R.: Petrology and Knowledge of the Earth's Crust.

5. Academician M. A. Kashkay of the Azerbaydzhanian Academy of Sciences and V. I. Aliyev: Structure and Composition of the Latest Yardymlin Iron Meteorite Fallen in Armenia, in 1959.

6. A. A. Gabrielyan, Corresponding Member, Armenian Academy of Sciences: Tectonic Differentiation of the Lesser Caucasus and Its Position in the Alpine Zone of the Southern U. S. S. R. and Adjacent Countries.

7. R. K. Tvaltvadze, Candidate in Physical and Mathematical Sciences, Structure of the Caucasian Segment of the Earth's Crust.

8. Academician A. P. Vinogradov. On the Origin of the Earth's Crust.

9. Academician Sh. A. Azizbekov, of the Azerbaydzhanian Academy of Sciences: The Araks Tectonic Zone (in Azerbaydzhani, Armenia, and Iran).

10. Academician S. S. Mkrtchyan of the Armenian Academy of Sciences: Development of the Mineral Resources of Armenia In the Last 40 Years.

11. E. M. Shikhlin'skiy, Candidate, Georgian Science: Climatic Differentiation of Mountain Countries.

The work of the Session reflected the substantial achievements of Armenian Scientists in the 40 years of the Soviet Armenia; it also demonstrated the unity of Soviet Science — an important factor in its future development.

CHRONICLE

ERRATUM

The title of A. I. Moskvitin's article in the February 1960 issue, page 32, should have read
"CLIMATIC DATA DETERMINING THE LOWER STRATIGRAPHIC BOUNDARY OF THE PLEISTOCENE."